JOURNAL OF ENVIRONMENTAL GEOGRAPHY
Journal of Environmental Geography 7 (1–2),1–9.
DOI: 10.2478/jengeo-2014-0001
ISSN: 2060-467X

18TH-CENTURY DAILY MEASUREMENTS AND WEATHER OBSERVATIONS IN THE SE-CARPATHIAN BASIN: A PRELIMINARY ANALYSIS OF THE TIMIŞOARA SERIES (1780-1803)

Ildikó Csernus-Molnár[1], Andrea Kiss[2], Edit Pócsik[3]

[1]Department of History, Central European University, Nádor u. 9, H-1051 Budapest, Hungary
[2]Institute of Hydraulic Engineering and Water Resources Management, Vienna University of Technology, Karsplatz 13, 1040 Vienna, Austria
[3]Institute of Communication in Foreign Languages, University of Szeged, Honvéd tér 6, H-6722 Szeged, Hungary
*Corresponding author, e-mail:kiss@hydro.tuwien.ac.at

Research article, received 26 January 2014, accepted 15 March 2014

Abstract

Covering a period of 23 years, the Timişoara (in historical Banat region; today SW-Romania) series is the earliest known long-term 18th-century daily measurement (temperature, pressure) and weather observation series (precipitation, sky coverage, meteorological extremes), preserved in the south-eastern lowlands of the Carpathian Basin. Based on data derived from the original weather diary of the royal pharmacist Karl Joseph Klapka, in this paper the early instrumental measurement and daily observation series is presented referring to the temperature, pressure, precipitation conditions, cloudiness, wind, types of precipitation and extreme weather events that occurred in Timişoara in the period of 1780-1803. The two daily temperature measurement series show very high (over r=0.95) correlations, while pressure series are also in good agreement with other known late 18th-century measurement series of the same period in the Carpathian Basin (Buda, Miskolc and Kežmarok). The Timişoara-series also contains important information concerning such weather extremes as the severe winter of 1784 or the unusual number of summer fog events in 1783 (presumably related to the Icelandic Lakagígar eruption), which are also reported in the present paper.

Keywords: historical climatology, early instrumental measurements, daily weather observations extreme weather events, 18th century

EARLY INSTRUMENTAL MEASUREMENTS AND METEOROLOGICAL OBSERVATIONS IN HISTORICAL HUNGARY

The 18th century is of basic importance from the viewpoints of the implementation of a uniform European measurement network and also from the development of meteorology and climatology in historical Hungary that covered the majority of the Carpathian Basin. The instrumental measurements and daily visual weather observations, carried out at the end of the 18th century, already provide much more accurate information on the weather conditions of the period – mainly on the temperature, pressure and precipitation events – than the earlier ones (e.g. Szőkefalvi-Nagy and Zétényi, 1965; Réthly, 1970). Both inside and outside of the *Societas Meteorologica Palatina* network, these records were usually made by (highly) educated people; in most cases by physicians, pharmacists, astronomers or members of the clergy (Brázdil et al., 2002, 2005; Demarée et al., 2002). Measurements and observations were initiated mainly on the basis of uniform methods, and thus, clear similarities in practices can be traced, for instance, in the Spanish, Swedish, Belgian, Italian and Czech analyses. These previous investigations can help in the methodology to process manuscripts, and correct as well as analyze data gained from the manuscripts (Barriendos et al., 2002; Brázdil et al., 2002, 2003; Camuffo, 2002a etc.).

Mainly due to the assembling work of Antal Réthly (1970), there are four known instrumental measurement series, combined with daily weather records, extending over several years from this time-period in the area of historical Hungary (*Fig. 1*). Firstly, the series of Buda station (Buda observatory) should be mentioned, which operated as a member of the *Societas Meteorologica Palatina* European measurement network, and its measurements were systematically documented in the Societas yearbooks until 1792. The temperature data and the daily visual observations from Miskolc (Hungary) between 1780 and 1801 (together with pressure series in 1780 and in the period 1794 - 1800), appeared in published form in 1794 and 1802 (by Benkö, 1794, 1802), and were partly analyzed (see Szőkefalvi-Nagy and Zétényi, 1965). Additionally, observations and measurements carried out in Kežmarok (in Slovakia; historical Késmárk) between 1789 and 1800 by János Genersich, remained in handwritten form (National Széchényi Library, Budapest, Fol. Germ. 114), similar to the data of Timişoara (in Romania; Temesvár in historical Hungary).

Fig. 1 Location of places mentioned in the paper (thick black line: 18th century borderlines of countries that belonged to the Hungarian crown; thin black line: present day country borders)

Recently, a complex processing of the Miskolc and Kežmarok originals has been carried out. These additional data can help us to provide an even more detailed picture of the weather patterns characterizing the Carpathian Basin in the late 18th century. This paper presents the measurements and daily records carried out in Timişoara: the digitalization of Klapka's written records – extending over 23 years – made it possible for us to perform the statistical analyses of the temperature and pressure data, as well as that of the daily meteorological observations. The importance of this work arises from the fact that, except for one publication concerning the Miskolc temperature series (Szőkefalvi-Nagy and Zétényi, 1965), the in-depth, detailed analysis of the Hungarian instrumental measurements and daily observations from the late 18th century, apart from the Buda(pest) series (the only late 18th century series included in the Societas Meteorologica Palatina network), have not been provided in published form.

THE KLAPKA-MANUSCRIPT AND DATA PROCESSING

The original manuscript can be found in its entirety in the Manuscript and Rare Book Collection of the Eötvös Loránd University Library in Budapest (catalogue No. E 40). The observer, Karl Joseph Klapka, was a royal pharmacist, coming from Znojmo (South-Moravia), who moved to Timişoara around the year 1780 (Szinnyei, 1900; Klapka, 1886). Although no information is currently available on the exact location of his measurements in Timişoara, it is quite probable that the measurements took place at the building (yard?) of the Pharmacy located in the city centre. Except for one or two missing days, an essentially continuous data series is available concerning the years between 1780 and 1803. The diary consists of data measured three times every day. The temperature measurements were carried out simultaneously by two thermometers (in Réaumur and Fahrenheit). In case of pressure measurements, the scale was divided according to the Paris measure in inches ('Grad') with values of 26, 27 and 28, every inch with 12 lines ('Dig') and every line with 10 units. In addition, Klapka regularly (three times a day) added information about the state of sky, cloudiness, wind variety and precipitation events (with types mentioned).

As the first step of digitalization, the numerical and textual information, available in the Klapka manuscript, was converted (i.e. typed in) to an excel file: in case of pressure the inch values had to be completed, while negative temperature values (with the sign "under zero") were converted to negative values. Due to the very uniform use of ink and the homogeneity of handwriting, we presume that the available manuscript was the copied, finalized version of the measurements. In a few cases, when a copying mistake of values was obvious (e.g. missing sign of negative value), we have corrected the actual number. Concerning the description of daily weather conditions, the precipitation-related narrative information was interpreted and transformed into numerical values (see details later). After digitalization and interpretation of the manuscript, the temperature and pressure series, and the daily records (written in Latin) were suitable for statistical analysis.

One of the great advantages of the Klapka-manuscript is that it provides continuous information for the period between 1780 and 1803, thus a data series characterized by unique continuity is available for this region, and for the south-eastern region of the Carpathian Basin in particular. It is an essentially important fact, since regular daily reports of the *Societas Meteorologica Palatina* were put aside for a few years after 1792. As a consequence, the daily observations carried out at the end of the century in Buda are not available; only monthly averages are known. The other advantage of this measurement series is that it originates from the last decades of the 18th century thus coincides with the so-called 'Maldá-anomaly', a period famous for its extreme weather conditions (Barriendos et al., 2003; Brázdil et al., 2003).

TEMPERATURE MEASUREMENTS

The early liquid-in-gas thermometers were very different from each other in many respects, namely deformation of the glass tube and bulb, polymerization of organic thermometer liquids, slipping of the scale or the capillary etc. which can cause measuring errors (Camuffo, 2002b). In this way, in order to be able to carry out statistical analysis of the data the two types of temperature measurement series provided by Klapka, had to be converted into one common unit of measure – in our case the Celsius degree scale. Furthermore, the comparison of the data recorded in the manuscript with the contemporary measurements of Buda, Kežmarok and Miskolc allows us to carry out a comprehensive analysis. In case of the Timişoara series, using the Celsius values derived from the measurements in Fahrenheit and Réaumur, morning, noon, evening and daily averages could be calculated. For the calculation of daily averages the Kämtz method was applied: (1* morning+1* noon+2* evening)/4 (see e.g. Dall'Amico and Hornsteiner, 2006; see also Kern, 2009).

The values measured by two different types of thermometers were practically identical to each other (r=1), correlated well both with the contemporary monthly temperature data series of Buda (r=0.98) (*Fig. 2*) and with the temperature data series of the period 1961-1990 measured in Timişoara (r=0.99) as well. Moreover, monthly values show strong correlation with the available contemporary measurement series of Miskolc (r=0.94) and Kežmarok (r=0.97). The second figure (*Fig.* 2) shows the monthly averages of temperature values in Timişoara and in Buda regarding the studied 23 years (1780-1803). It is conspicuous that, after 1783, there is a difference of approximately 3 degrees on average between the two observation sites: the monthly temperature values in Timişoara are all-time higher than those of Buda (which fits the general differences in climate), which is particularly visible in case of mild winters such as 1786, 1792, 1796, 1798 or 1800. It has to be emphasized, however, that difference between the Buda and Timişoara temperature series is much less significant in case of summer, and even less detectable in case of autumn and spring values (see *Fig.* 2). Moreover, the daily course of mean temperatures referring to the entire study period, in general, reflects values comparable to the 20th-century differences in climate (*Fig. 3*). Nevertheless, the generally high winter values and relatively small differences in the course of daily temperature might allow the possibility of some influence of heating (it was a common practice /especially in towns/ to put the thermometers on the wall, or place them in courtyards etc.).

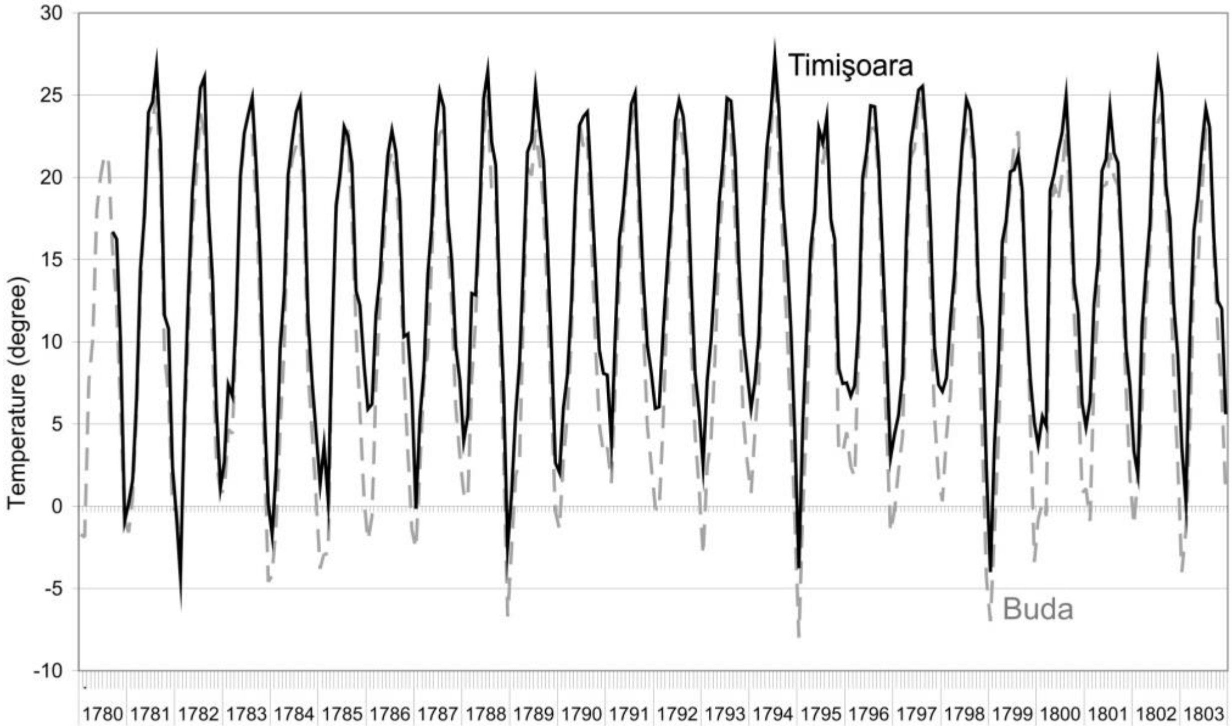

Fig. 2 Monthly averages of temperature values measured in Timişoara and Buda in 1780-1803 (ºC)

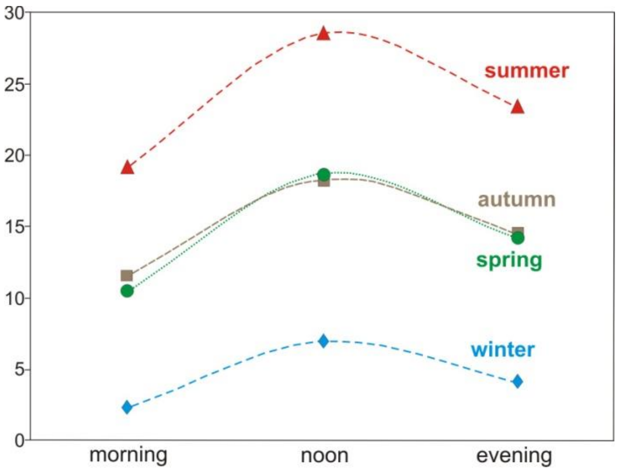

Fig.3 Daily course of temperature measured in 1780-1803, in Timişoara (ºC)

The values transformed are still raw data series, and it follows from that, that a deeper and more accurate analysis should be carried out in future. Comparing the annual courses of temperature series in 1780-1803 and 1961-1990 (*Fig. 4*), it comes clear that monthly temperatures measured in 1780-1803 are generally 2-3 ºC higher than those measured in 1961-1990 in Timişoara; this difference can reach 5 ºC in winter months. Thus, higher degrees can be systematically detected not only compared to the contemporary Buda measurements (located ca. 250 km northwest to Timişoara with the Great Hungarian Plain in between), but also to the present-day (1961-1990) Timişoara data series. What could be the reason of this difference, especially in winter time? The clear reasons are unknown; nevertheless, the urban location of the pharmacy, and the contemporary practices of settling the instruments (maybe an inner yard location, thermometer placed on the wall) or some technical reasons (installation) can also be responsible for the differences. Since positive temperature anomalies can primarily be detected during mild winters and they are less visible during summers, the effect of insolation seems to be a less likely reason for the higher temperature values.

Based on the Klapka-measurements, in Timişoara especially the summers of 1781, 1788, 1794 and 1802 turned out to be outstanding warm. The extraordinary hot summer of 1783, which was reported in large parts of Central Europe (Brázdil et al., 2003), was not an unusually hot summer in Timişoara concerning monthly averages (see *Fig. 2*). On the basis of monthly average temperatures, the hardest winters in Timişoara occurred in 1782, 1789, 1795 and 1799, which were otherwise accompanied by ice jam floods along the Danube in the Carpathian Basin, similarly to some other areas in Central Europe (see e.g. Strömmer, 2003; Kiss, 2007; Glaser, 2008; Brázdil et al., 2010).

Due to its severe conditions and the great and extraordinary winter- and spring-flood events, the winter of 1784 (after the great Lakagígar eruption in Iceland) is considered as one of the most famous European extremes in the last decades of the 18th century (for a European overview of flood waves, background and consequences, see Brázdil et al., 2008). Although the winter of 1784 was clearly colder than usual in Timişoara, if we merely take the temperature averages into consideration, it was not one of the coldest winters of the study period. It is rather due to the great temperature variations (very cold periods were interrupted by rather mild ones), reflected in the course of daily temperatures as well as the values of daily maximum and minimum temperatures (*Fig. 5*), that this winter was reported as a notably cold one in the Carpathian Basin, when both ice jam flood (e.g. in December on the Maros river) as well as floods due to snow-melt (from January) occurred around Timişoara and in the Banat area (see e.g. Kiss et al., 2006; Kiss and Csernus-Molnár, 2008; Kiss et al., 2008).

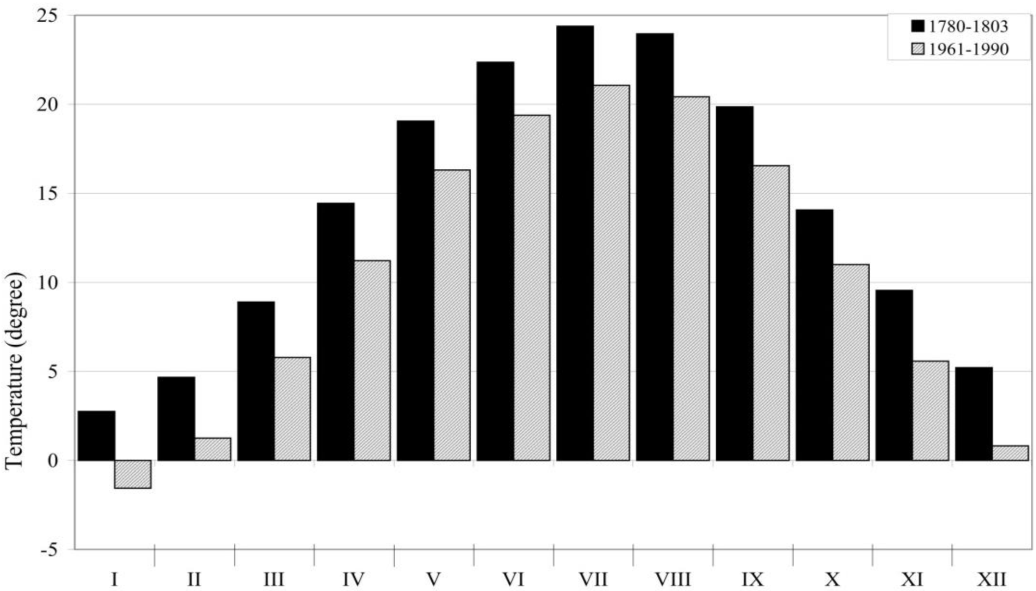

Fig.4 Annual course of measured temperature values in Timişoara: 1780-1803, 1961-1990 (in ºC)

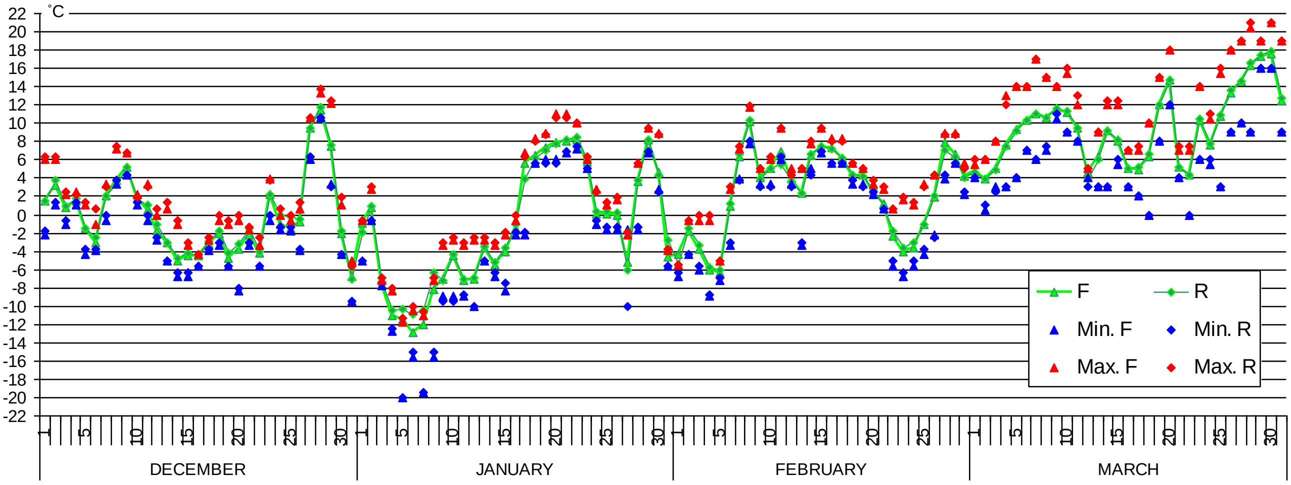

Fig. 5 Daily measured temperature (provided in ºC; thermometers: F=Fahrenheit; R=Réaumur): average (green lines), maximum (red) and minimum (blue) values in Timişoara between 1 Dec. 1783 and 31 March 1784

The late 18th century was also rather rich in individual temperature extremes in Central Europe, and also in Timişoara and the Carpathian Basin. In Timişoara, the coldest December occurred in 1788 (-2.6 ºC measured by Klapka; in 1961-1990: 0.8 ºC), the coldest March (0.3 ºC; 1961-1990: 5.8 ºC) and spring were measured in 1785 (1.7 ºC; 1961-1990: 11.1 ºC), whereas the warmest April occurred in 1800 (19.2 ºC; 1961-1990: 11.2 ºC).

AIR PRESSURE MEASUREMENTS

Similar to temperature, pressure measurements were provided three times per day in a rather systematic way. Even if Klapka provided no evidence on the measurement unit of his pressure data series, based on the values included in the manuscript, the pressure data presumably were recorded in Paris inches, which was widely applied in this period in Europe. From the original daily values, transformed to hPa form, daily and monthly averages were calculated (*Fig. 6*): although the annual pressure averages show some parallels to those measured in Timişoara between 1961 and 1990, the non-corrected values are significantly lower than the ones measured in the 20th century. Apart from the fact that these are still raw values (not corrected to temperature or reduced to sea elevation), the values will change after corrections, the problem of low values clearly needs further elaboration.

Similarly to the analysis of the temperature data, for further quality assessment of the pressure measurements carried out in Timişoara, the daily data series of Buda were applied, the raw data of which were obtained from the Mannheim series of the *Societas Meteorologica Palatina* (*Ephemerides… 1781-1792*), and transformed to hPa using the aforementioned formula. As, similarly to the

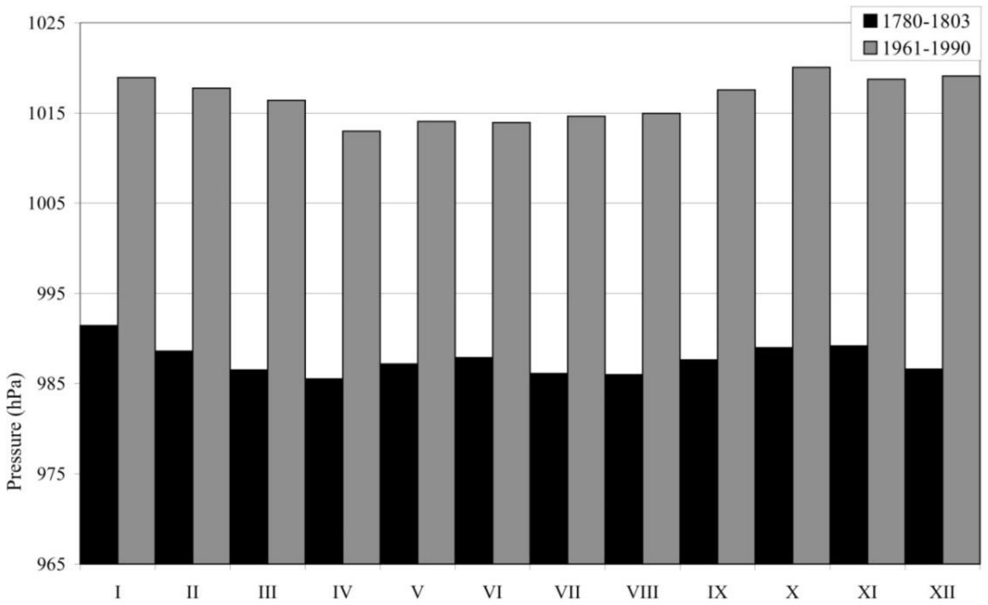

Fig. 6 Comparison of monthly averages of pressure values (without corrections), measured in the periods 1780-1803 and 1961-1990 in Timişoara

temperature series, the pressure series from Buda also contains values measured three times every day, it could be compared with the pressure values measured in Timişoara. The correlation between these monthly averages is also strong (r=0.58; <0.01 significance level): the strongest correlation can be detected in winters (e.g. in January of 1784: r=0.98), while the lowest were in summers (e.g. in July of 1783: r=0.37). The daily values are also comparable with measurements of Miskolc (r=0.24; 0.01 significance level) and Kežmarok (r=0.87; <0.01 significance level; with rel. short overlapping period). Low correlations with the Miskolc pressure measurement series can also be caused by the fact that the Miskolc series are rather problematic as (when provided) only one value is available for each day without any information on the time of measurement (see also Szőkefalvi-Nagy and Zétényi, 1965). Comparing the Timişoara (raw) pressure series (1780-1803) to the measured (corrected) values of 1961-1990, similarly strong connection can be detected (r=0.65; <0.01 significance level). This fact also highlights the considerable potentials of this early and almost uninterruptedly recorded 23-year long series.

DAILY OBSERVATIONS, WITH SPECIAL EMPHASIS ON PRECIPITATION PATTERNS

Beyond recording the measured temperature and pressure values, Klapka also provided a systematic description of weather events (e.g. precipitation, atmospheric phenomena) three times every day in Latin. Since the application of contemporary terminology, the use of specific words have special importance in proper understanding of weather conditions, it is necessary to provide the precise, present-day interpretation of the more important terms. Terms appearing most often in the manuscript refer to transparency, cloudiness, type of precipitation (e.g. snow, fog), motion of air (e.g. windy) or to temperature (e.g. cold, hot). However, some individual phenomena (e.g. rainbow) also appear in the manuscript (see *Table 1*).

In addition to the measured temperature and pressure values, the aforementioned observations provide us more information for instance about the number of days with precipitation, the physical condition of precipitation, the number of cloudy or of foggy days. From all these information the systematically recorded daily evidence on precipitation might be the most important.

PRECIPITATION OBSERVATIONS

A significant further section of the daily observation records refers to precipitation. Although no precipitation measurements were carried out, Klapka recorded the information on precipitation every day. From this evidence, it is possible to calculate days with precipitation, dividing the information according to precipitation types. After aggregating the number of days with precipitation summed up for each month, the numbers of rainy, snowy and mixed days were separately grouped (*Fig. 7*). At the same time the manuscript also provides some information on trace precipitation (mostly about the number of foggy days - see later), which were treated separately from normal precipitation information. We have to note that even if some comparisons were carried out with precipitation measurement data series, the number of days with precipitation is not equivalent to the amount of precipitation measured in other places. Moreover, it is possible that precipitation events during the night were sometimes not recorded in the manuscript.

Precipitation results (quantity) were compared with the daily precipitation data of Buda, published in the volumes of the *Societas Meteorologica Palatina* (*Ephemerides… 1781-1803*). The correlation between the Buda and Timişoara datasets shows connection on 5% significance level (r=0.27). Thus, despite the great distance (over 250 km) and the differences in precipitation patterns, some connections could be detected with the contemporary measured precipitation datasets of Buda, which supports the further potentials of the late 18th-century daily precipitation information recorded in Timişoara.

At the same time, the manuscript also reflects on some of the dry (such as 1794, 1797, 1800, 1802 etc.) and rainy years (such as 1786, 1795, 1801). In the latter case,

Table 1 Basic terminology applied in the descriptions of daily visual observations

Latin term	**English meaning**	**Latin term**	**English meaning**
1. Sky coverage, transparency		3. Motion of air	
nubilum, nubiles	cloudy, clouds	*ventus, ventosus*	wind, windy
(*semi-*) *serenum* (*sereno*)	(half-) clear sky (air)		
Obscurum	unclear, grey/dark weather (sky)	4. Temperature-related terms	
2. Types of precipitation, physical condition		*frigidus, frigeo* (*frigefacto*)	cold, cooling down
pluvia (*pluvius, pluviosus*)	rain (rainy)	*calidus* (*caleo*)	Warm
nives, ningo/*ninxi*, (*nivosus*)	snow, snowing (snowy)	5. Atmospheric phenomenon	
Nebula	fog	*arcus pluvius*	Rainbow

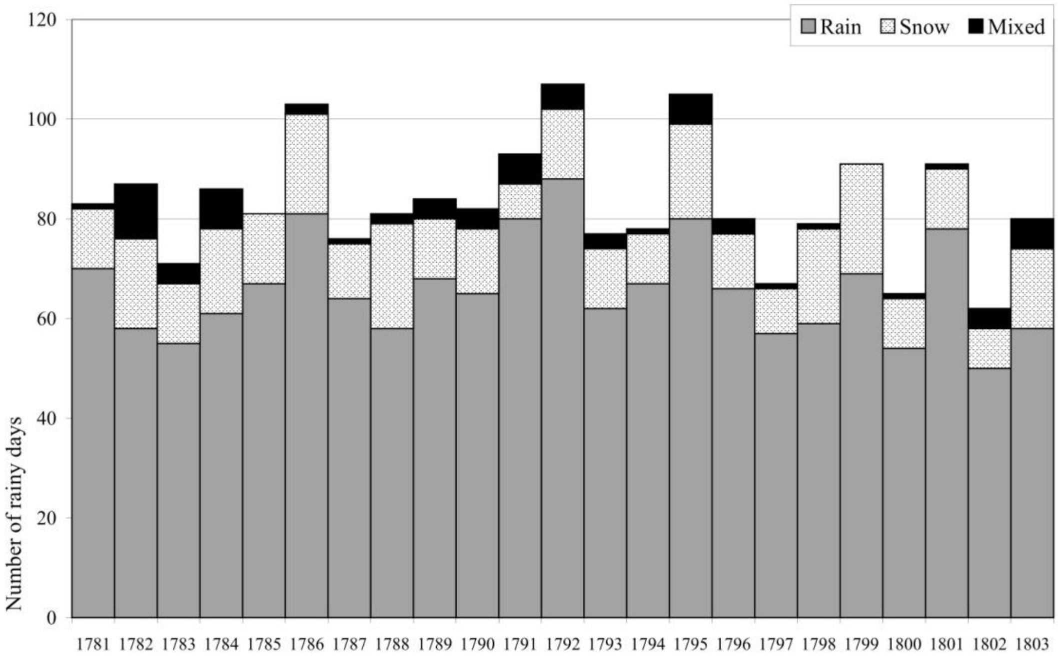

Fig.7 Number of days with precipitation per year in Timişoara (trace precipitation excluded): 1781-1803

the number of days with more significant precipitation events exceeds 100 (e.g. 1786, 1792, 1795; see *Fig. 7*). Concerning anomalous and extreme months, the springs of 1794, 1801 and 1802, the summers of 1784, 1788, 1793, 1800, 1802 and 1803, the autumns of 1783, 1793, 1797 and 1802, the winters of 1790, 1794, 1796 and 1801 appeared to be particularly dry. Nevertheless, the summer of 1801, the autumns of 1791, 1794 and 1801, the winters of 1781, 1792 and 1795 seem to be particularly rainy. On an annual scale 1790, 1793, 1802 (where not only the spring but also the autumn) and 1803 were dry - at least based on the number of days with precipitation.

As we pointed out earlier, it is an important fact that the number of days with precipitation events is not always closely related to the amount of precipitation: that is why, for example, the otherwise famous drought year of 1794, does not appear as a great singularity in the record. Moreover, the (inter)annual distribution of days with precipitation is also an important, further question to be discussed, greatly influencing drought patterns (especially in case of agricultural drought). The distribution of days with precipitation, presented annually, still provides some rather interesting information, even concerning the drought year of 1794. For instance, 77 days with precipitation events were recorded in the manuscript concerning that year, but only 12 days were noted to be rainy in the entire spring, and no precipitation at all was recorded in March. Moreover, there were no more than 4 rainy days in August. Reports from other documentary evidence, such as contemporary newspapers (e.g. in *Magyar Hírmondó* 19 August 1794), are also available about the great drought ('beyond any record') of spring and summer on the Great Hungarian Plain which - among other evidence - provide a rather detailed and clear picture about the severe lack of precipitation mainly in spring and summer, but practically also in winter and in the preceding years (for a short overview of scientific literature about this drought event see Kiss, 2009).

Another interesting example can be raised from 1781: the annual distribution of days with precipitation events was rather extreme during the year. Whereas the winter (1781) was remarkably rainy, the spring and the summer were rather dry. The year of 1792 was similar. Despite the fact that Klapka observed more than 100 days with notable precipitation, the highest number of days with precipitation events was observed during winter, while spring and summer were dry.

TRACE PRECIPITATION

Although similarly to precipitation evidence, trace precipitation was probably not always recorded in time, some conclusions can be drawn, especially concerning the number of foggy days provided in the manuscript. Probably the most interesting example is the extraordinary event which reflects from the written records occurred in this period: an unusually large number of days (14) with fog was recorded in Timişoara during the summer of 1783. This is rather extreme in this season and shows an unambiguous anomaly compared with the other summers of the studied period (1781-1803); the number of foggy days in summer was usually between 2 or 3, but never (apart from 1783 summer) exceeded 7 (*Fig. 8*). The unusually large number of foggy days is in good agreement with the great number of dry fogs recorded all over Europe in this summer (also in the actual *Societas Meteorologica Palatina* volume), which was one of the 'side-effects', impacts of the Lakagígar eruption in Iceland during the summer of 1783.

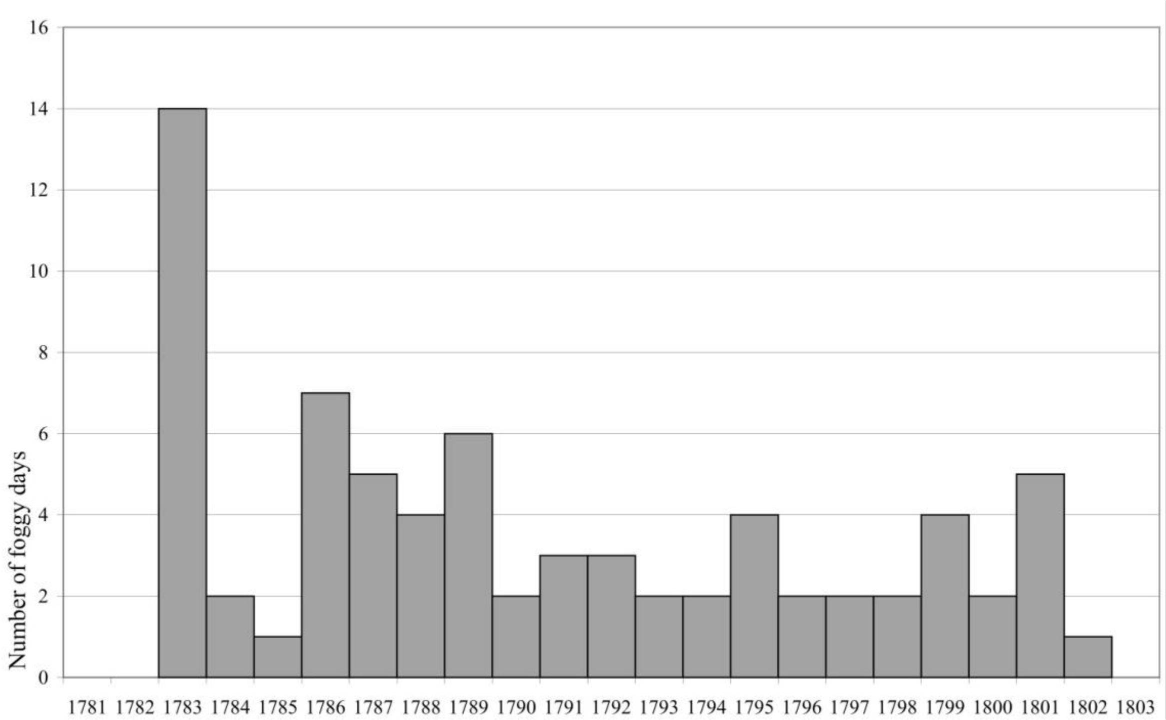

Fig. 8 Number of foggy days in summer (1781-1803), recorded in the Klapka-manuscript

The consequences of this major volcanic eruption (lasted for 6 months) have been extensively presented in many studies, including atmospheric and weather impacts, the extreme precipitation and temperature conditions (hot summer with numerous thunderstorms, extreme winter etc.) together with related flood waves and ice jam floods during the hard winter of 1784 in Europe (see e.g. Stothers, 1996; Self and Rampino, 1998; for Hungary: Kiss et al., 2006; latest European overview: Brázdil et al., 2010). Apart from other consequences, a large number of contemporary reports can also be found concerning the dry fog phenomenon in the Carpathian Basin: newspapers, such as the *Magyar Hírmondó* (12. July 1783.) or the *Pressburger Zeitung* (30. July 1783.) published reports about this unusual but rather interesting phenomenon.

CONCLUSIONS AND OUTLOOK

The Timişoara weather diary written between 1780 and 1803 by K. J. Klapka (pharmacist) represents the south-easternmost meteorological instrumental measurements in the Carpathian Basin from the late 18th century. The importance of the Timişoara measurements and daily observations lies in their continuity (for 23 years, with only some days missing), construction and accuracy, and in the fact that they date back to an important period of historical climatology, namely the second part of the 18th century (the so-called 'Maldá-anomaly') which was famous for its temperature and precipitation extremes. Due to the fact that the *Societas Meteorologica Palatina* volumes were not published after 1792 and the daily measurements at Buda are not available for several years after 1792 (only monthly sums are known), the measurement series of Timişoara represents the single, continuous daily temperature and pressure data series from this region known at present (the series of Miskolc and Kežmarok are incomplete, and with some further quality problems; see Csernus-Molnár and Kiss, 2011).

After the digitalization of the measurement series of Timişoara, the measured data could be compared to the results of the contemporary measurements carried out in Buda, Miskolc and Kežmarok: strong correlations were detected between both the temperature and pressure series, which prove the further elaboration potentials of the datasets included in the manuscript. Concerning the daily weather records the information on precipitation play an important role as the analysis of the number of days with precipitation provided some further parallels to the contemporary Buda measurement series. Moreover, the data on trace precipitation could provide some interesting information concerning the large number of foggy days observed in the summer of 1783, which might be in relation with the consequences of a major Icelandic (Lakagígar) volcanic eruption in 1783.

References

Barriendos, M., Martin-Vide, J., Pena, J. C., Rodriguez, R. 2002. Daily Meteorological Observation in Cádiz – San Fernando. Analysis of the Documentary Sources and the Instrumental Data Contect (1786-1996). *Climatic Change* 53, 151–170. DOI: 10.1007/978-94-010-0371-1_6

Barriendos, M. – Llasat, M. C. 2003. The Case of the 'Maldá' anomaly in the Western Mediterranean Basin (AD 1760-1800): An Example of a Strong Climatic Variability. *Climatic Change* 61, 191–216. DOI: 10.1023/a:1026327613698

Benkö, S. 1794. Ephemerides Meteorologico-Medicae annorum 1780…..1793. Vindobonae: Typis Alb. Ant. Patzowsky. Tom. I-V. 258, 330, 283, 323, 164 p.

Benkö, S. 1802. Novae Ephemerides astronomico-medicae annorum 1794…..1801. Vindobonae: Typis Alb. Ant. Patzowsky. 204 p.

Brázdil, R., Valášek, H., Sviták, Z., Macková, J. 2002. History of weather and climate in the Czech Lands. Vol. V: Instrumental Meteorological Measurements in Moravia up to the End of the Eighteenth Century. Brno: Masaryk University. 250 p.

Brázdil, R., Valášek, H. , Macková, J. 2003. Climate in the Czeh Lands the 1780s in Light of the Daily Weather Records of Parson Karel Bernard Hein of Hodonice (Southwestern Moravia): Comparison of Documentary and Instrumental Data. *Climatic Change* 60, 297–327. DOI: 10.1023/a:1026045902062

Brázdil, R., Pfister, C., Wanner, H., van Storch, H., Luterbacher, J. 2005. Historical Climatology in Europe , The State of the Art. *Climatic Change* 70, 363-430. DOI: 10.1007/s10584-005-5924-1

Brázdil, R. , Kiss, A. , Luterbacher, J. , Valášek, H. 2008. Weather Patterns in Eastern Slovakia 1717-1730, Based on Records from the Bresslau Meteorological Network. *International Journal of Climatology* 28/12, 1639–1651. DOI: 10.1002/joc.1667

Brázdil, R., Demarée, D.R., Deutsch, M., Garnier, E., Kiss, A., Luterbacher, J., Macdonald, N., Rohr, Ch., Dobrovolný, P., Kolář, P., Chromá, K. 2010. European floods during the winter 1783/84: scenarios of an extreme event during the 'Little Ice Age'. *Theoretical applied Climatology* 100, 163–189. DOI: 10.1007/s00704-009-0170-5

Camuffo, D. 2002a. History of the long series of daily air temperature in Padova (1725-1998). *Climatic Change* 53, 7–75. DOI: 10.1007/978-94-010-0371-1_2

Camuffo, D. 2002b. Calibration and Instrumental Errors in early Measurements of Air Temperature. *Climatic Change* 53, 279–329. DOI: 10.1007/978-94-010-0371-1_11

Csernus-Molnár, I. , Kiss, A. 2011. A XVIII. század végi magyarországi műszeres mérések feldolgozási és vizsgálati lehetőségei (Research and study possibilities of late 18th-century instrumental weather measurement series in Hungary). In: Kázmér, M. (ed.). *Környezettörténet 2* (Environmental history 2). Budapest: Hantken Kiadó. 203–214.

Dall'Amico, M., Hornsteiner, M. 2006. A simple method for estimating daily and monthly mean temperatures from daily minima and maxima. *International Journal of Climatology* 26, 1929–1936. DOI: 10.1002/joc.1363

Demarée, G. R., Lachaert, P. J., Verhoeve, T., Thoen, E. 2002. The Long-Term Daily Central Belgium Temperature (CBT) Series (1767-1998) and Early Instrumental Meteorological Observations in Belgium. *Climatic Change* 53, 269–293. DOI: 10.1007/978-94-010-0371-1_10

Ephemerides Societatis Meteorologicae Palatinae Observationes. Mannheim: Ex Officina Novae Societatis Typographicae. Tom. I-XII, 1781-1792.

Genersich, J. 1789-1800. Meteorologische Beobachtungen in dem Jahr 1789. Országos Széchényi Könyvtár, Kézirattár, Fol. Germ. 114.

Glaser, R. 2008. Klimageschichte Mitteleuropas. 1200 Jahre Wetter, Klima, Katastrophen. Darmstadt: Primus Verlag. 264 p.

Kern, Z., Popa, I. 2009. Assessing temperature signal in X-ray densitometric data of Norway spruce and the earliest instrumental record from the Southern Carpathians. *Journal of Environmental Geography* 2(3-4), 15–22.

Kiss, A., Sümeghy, Z., Danku, Gy. 2006. Az 1783-1784. évi szélsőséges tél és a Maros jeges árvize (Severe winter of 1783-1784 and the ice flood on the Maros river). In: Kiss, A., Mezősi, G., Sümeghy, Z. (eds.). Táj, környezet és társadalom / Landscape, Environment and Society. Szeged: SZTE. 353–362.

Kiss, A., Sümeghy, A., Fehér, Z. Zs. 2008. A Maros 18. századi áradásai és egy jellemző téli árvizének területi hatásai (18th-century floods of the Maros river and the areal consequences of a winter flood). In Füleky, Gy. (ed.) A táj változásai a Kárpát-medencében. Az erdélyi táj változásai(Landscape changes in the Carpathian Basin. Changes of the Transylvanian landscape). Gödöllő: Szent István Egyetem. 94–100.

Kiss, A. 2007. „Suburbia autem maxima sui in parte videntur prorsus esse deleta" , Danube icefloods and the pitfalls of urban planning: Pest and its suburbs in 1768-1799. In: Kovács, Cs. (ed.) *From Villages to Cyberspace.* Szeged: University Press. 271–282.

Kiss, A. 2009. Historical climatology in Hungary: Role of documentary evidence in the study of past climates and hydrometeorological extremes. *Időjárás* 113(4), 315–339.

Kiss, A. , Csernus-Molnár, I. 2008. Időjárási viszonyokhoz kapcsolható szélsőségek területi vonatkozásai a Temesi Bánságban: 1780-1800 (Areal consequences of weather-related extremes in the Temesi Bánság/Banat area: 1780-1800). In Füleky, Gy. (ed.) A táj változásai a Kárpát-medencében. Az *erdélyi táj változásai* (Landscape changes in the Carpathian Basin. Changes of the Transylvanian landscape). Gödöllő: Környezetkímélő Agrokémiáért Alapítvány-Szent István Egyetem. 101–106.

Kiss, A., Sümeghy, Z., Danku, Gy. 2006. Az 1783-1784. évi szélsőséges tél és a Maros jeges árvize (The severe winter of 1783-1784 and the ice-flood of the Maros river). In Kiss, A., Mezősi, G., Sümeghy, Z. (eds.) Táj, környezet és társadalom / Landscape, Environment and Society. Szeged: University Press. 353–362.

Klapka, C.J. 1803. Observationes Thermometricae et Barometricae a 1a Septembris 1780. usque ultimam Decembris 1803. Temesvarini factae per C. J. Klapka. Eötvös Loránd University Library, Manuscript and Rare Book Collection, Cat. number E 40.

Klapka, Gy. 1886. Emlékeimből (From my memories). Budapest: Franklin Társulat. 4–5.

Réthly, A. 1970. Időjárási események és elemi csapások Magyarországon 1701-1800-ig (Weather events and natural disasters in Hungary in 1701-1800). Budapest: Akadémiai Kiadó. 519-565.

Self, S., Rampino, M.R. 1998. The Relationship between Volcanic Eruptions and Climatic Change: Still a Conundrum. *EOS* 69/6, 74–86.

Stothers, R.B. 1996. The great dry fog of 1783. *Climatic Change* 32, 79–89. DOI: 10.1007/bf00141279

Strömmer, E. 2003. Klima-Geschichte. Methoden der Rekonstruktion undhistorische Perspektive. Ostösterreich 1700 bis 1830. Forschungen und Beiträge zur Wiener Stadtgeschichte 39. Vienna: Deuticke. p. 325.

Szinnyei, J. 1900. Magyar írók élete és munkái (Life and works of Hungarian writers). Budapest: Hornyánszky Viktor. Vol. 6. 466.

Szőkefalvi-Nagy, Z., Zétényi, E. 1965. Egy XVIII. századi magyar meteorológus: Benkő Sámuel miskolci orvos (An 18th-century Hungarian meteorologist: Sámuel Benkő, the doctor of Miskolc). *Borsodi Szemle* 9, 51–56.

JOURNAL OF ENVIRONMENTAL GEOGRAPHY
Journal of Environmental Geography 7 (1–2), 11–22.
DOI: 10.2478/jengeo-2014-0002
ISSN: 2060-467X

DATASET FOR CREATING PEDOTRANSFER FUNCTIONS TO ESTIMATE ORGANIC LIQUID RETENTION OF SOILS

Hilda Hernádi*, András Makó

Department of Crop Production and Soil Science, Georgikon Faculty, University of Pannonia, Deák F. u. 16, H-8360 Keszthely, Hungary
*Corresponding author, e-mail: hhilduci@gmail.com

Research article, received 27 January 2014, accepted 14 March 2014

Abstract

Soil properties characterising pressure-saturation relationships (P-S), such as the fluid retention values or the fitting parameter of retention curves are basic input parameters for simulating the behaviour and transport of nonaqueous phase liquids (NAPLs) in subsurface. Recent investigations have shown the limited applicability of the commonly used estimation methods for predicting NAPL retention values in environmental practice. Alternatively, building pedotransfer functions (PTFs) based on the easily measurable properties of soils might give more accurate and reliable results for estimating hydraulic properties of soils and enable the utilisation of the wide range of data incorporated in Hungarian and international datasets. In spite of the availability of several well-established PTFs to predict the water retention of soils only a limited amount of research has been done concerning the NAPL retention of soils. Thus, in our study, data from our recent NAPL and water retention measurements were collected into a dataset containing the basic soil properties as well. Relationships between basic soil properties and fluid retention of soils with water or an organic liquid (Dunasol 180/220) were investigated with principal component analysis. NAPL retention of soil samples were determined with PTFs, based on basic soil properties and their derived values, and using a scaling method. Result of the statistical analysis (SPSS 13.1) revealed that using PTFs could be a promising alternative and could give more accurate results compared to the scaling method both for determining the NAPL saturation or the volumetric NAPL retention values of soils.

Keywords: NAPL retention, pedotransfer function, hydraulic characteristics, Leverett equation

INTRODUCTION

From the 1980s increased attention has been received to improve understanding the relevant characteristics and processes in flow and transport of subsurface nonaqueous phase liquids (NAPL) firstly in petrol industry and later to help design remedial strategies. Knowledge of pressure-saturation (P-S) relationships is essential for simulating the fate and transport of nonaqueous phase liquids (NAPLs) in subsurface with any type of models. Nowadays, as the measurements of hydraulic parameters are costly and time consuming the development of accurate estimation method is preferred.

In environmental practice the NAPL retention of soils is commonly determined by the table of average empirical pressure-saturation (P-S) values (dePastrovitch et al., 1979), table of average fitting parameters of the van Genuchten equations proposed by Carsel and Parrish (1988) (e.g. in RETZ model – van Genuchten et al., 1991; HSSM model – Weawer et al., 1994), the different modified versions of the Leverett function (Leverett, 1941) or their combination. In addition, Beckett and Joy (2003) created a dataset of calculated fitting parameters based on the modified scaling method suggested by Lenhard and Parker (1987). Nevertheless, tables of average NAPL retention or fitting parameter values of fluid retention curves do not accurately represent the variability of soils with different physical and chemical properties. All type of the modified Leverett function is valid for ideal porous systems. These estimation methods do not take into account the different interactions between the various fluids and the porous media, thus their application might be limited for natural soils (i.e. well aggregated, higher organic matter or clay content, etc.) soils. Furthermore, they have not been properly validated (only a few methods validated with column experiments, but these were carried out in most cases with glass beads and/or sands) (Rathfelder and Abriola, 1996; Makó and Hernádi, 2013).

In the commonly used fluid retention measurement methods usually only the main drainage curves are measured. The more accurate determination of the main drainage curve of soil has key importance in determining the transport parameters (e.g. residual organic liquid content, penetration depth and time). In addition, main drainage curves could be the basis of determining the hysteretic and scanning curves in hysteretic models.

Predicting of the physical, chemical and biological properties of soils with pedotransfer functions (PTFs) is a fast-developing field and several well-established PTFs are available to predict the water retention of soils both in Hungary (Rajkai, 2004; Rajkai et al., 1996; 2004; Nemes, 2003; Makó et al., 2005 and Tóth, 2011) and abroad (Minasny et al., 1999; Wösten et al., 1999; Rawls et al., 2001, etc.). Most of these estimation methods have already been incorporated into numerical algorithms such as SOILPAR 2.0 (Acutis and Donatelli, 2003), Neuro Multistep (Minasny et al., 2004), TALAJTANonc 1.0 (Fodor and Rajkai, 2011) or the k-Nearest (Nemes et al., 2008), etc. However, only a few studies had begun focusing on creating PTFs concerning the soils organic liquid retention capacity.

Experiences in creating PTFs for water retentions may be essential tool for obtain the best possible estimation method for predicting the NAPL retention. PTFs for water retention (as response variable) may be created for the measured point values of the pressure saturation curves (point estimation) or for the fitting parameters of the hydraulic functions (parameter estimation) (Brooks and Corey, 1964; Brutsaert, 1966; van Genuchten, 1980) based on the easily measurable basic soil properties or their derived values (predictor variables) (Wösten, 1995).

The parameter estimation methods are widely used in environmental and soil hydrological practices because the simulation models mainly use these fitting parameters as input variable. Moreover, these parameters are equal to those of the equations predicting the soil hydraulic conductivity and relative permeability (van Genuchten, 1980; Lenhard and Parker, 1987; Chen et al., 1999). If the results of two point PTFs were to be compared or if no measured water retention points are available for a particular PTF, (or vice versa) the comparison of the retention curves with using the fitted hydraulic parameters (Minasny et al., 1999; Rawls et al., 2001) or calculating the estimated water contents at the desired pressure heads by linear interpolation were suggested (Tietje and Tapkenhinrichs 1993; Schaap and Leij, 2001).

For parametric methods it is common to predict logarithmic transformed values of α ($\ln \alpha$) and n ($\ln n-1$) to convert the distribution of the parameters into a more statistically normal distribution (e.g., Rawls and Brakensiek, 1985; Wösten et al, 1999). Both in case of water and NAPL retention estimation, application of similar soil properties (bulk density or texture class information, organic matter and carbonate content, etc.) or their inherited values (e.g. the averaged values of particle size data) were suggested as independent variables (Makó, 2004; Makó and Elek, 2006). In case of predicting the water retention the texture, morphology etc. are commonly used as a grouping factor in developing PTFs (Wösten et al. 1995; Schaap et al. 1999, 2001; Pachepsky and Rawls 2004). Statistical attributes of comparing PTFs for predicting water retention can potentially be adapted for the investigation of NAPL retention estimation methods. Recently, the determination of accuracy (with the working dataset), uncertainty and reliability (with test data) of the predictions, R, R2, mean error (ME), mean square error (MSE), root mean square error (RMSE), the unbiased root mean square errors (URMSE), etc. and their complete calculation are recommended, to acchieve a comprehensive verification (Pachepsky and Rawls, 2004). For the comparison of the accuracy of the fitted fluid retention curves the calculation of ZAPF values were suggested by Rajkai (2004). The AIC (Akaike Information Criterion) value offers the possibility to compare the efficiency of different estimation or fitting methods with various numbers of parameters and the models with lower error can be selected with Fisher's test (Rajkai, 2004). The uncertainty in input data can be evaluated using Monte Carlo analysis (Minasny et al. 1999), procedures based on fuzzy rules (McBratney et al., 2002) or with the Bootstrap Method (Carsel and Parish, 1988).

Nowadays, the development of inference systems (e.g. SINFER) to select the proper PTFs with minimum variance based on logical rules (McBratney et al., 2002)., or the application of data driven methods, e.g. support vector machines (Lamorsky et al., 2008) might be a challenging topic. The stability of the estimated coefficients can be investigated using the double cross-validation techniques (K-fold, Leave-one-out, Jackknife or Delete-d methods) with randomly split the data (Pachepsky and Rawls, 2004). According to Tóth et al., (2013) it may be sufficient to random spilt the data in proportion of 90:10 (working and test data). Besides, many authors suggested the detection of outliers for calibrating PTFs. To our current knowledge, there is only one research had begun for creating PTFs concerning the soils organic liquid retention in Europe (Makó, 1995; Makó, 2002; 2004). In spite of the large number of measurements to determine the NAPL retention of soils, databases from measured NAPL retention data have not been created until now. Beckett and Joy (2003) created a dataset of calculated fitting parameters based on the modified scaling method suggested by Lenhard and Parker (1987) and Parker et al. (1987) but this contains the scaled NAPL retention values calculated from the fitted water retention values of HYPRES database. Creating PTFs for NAPL-retention might be promising because a large amount of measured basic data is available in national and Hungarian databases (HYPRES, UNSODA v2.0, HUNSODA, EU HYDI, MARTHA etc.), which have already been used effectively in developing PTFs for estimating water retention (Wösten et al., 1995; Nemes et al., 2003, 2008; Makó et al., 2005; Tóth et al., 2006, 2013; Lilly, 2010). Moreover, the up-to-date hydrodynamic and transport models enable the adaptation of GIS (Geographic Information Systems) datasets (e.g. GMS - Groundwater Modelling System and Argus Open Numerical Environments - ARGUS ONE), which allow for cartographic representation with different commonly used software applications in the environmental engineering practice (such us SURFER-GRAPHER or AUTOCAD). In the 1990s a series of investigations for measuring NAPL retention of soils and mineral mixtures with the pressure plate method to create PTFs for organic liquid retention, in Hungary were started (Makó, 1995; Makó, 2002).

In this study a dataset from these recent measurements was created and analysed by statistical methods (SPSS 13.1). After the preliminary analysis (descriptive statistics and outlier detections), PTFs were built for predicting the fitting parameters of NAPL retention curves. Then, the NAPL retention of soils was predicted with classPTFs (for selected four texture groups) and using the scaling method of Lenhard and Parker (1987). Afterwards, the accuracy and reliability of predicting NAPL retention with different estimation methods were compared.

DESCRIPTION OF THE DATASET

The dataset contains the physical and chemical properties and the fluid retention data of five measurement series collected from 1991 to 2011 (*Table 1-2*). Dataset contains 369 disturbed and undisturbed samples, 40 soil profiles with 107 genetic layers and various types of soils (*Fig. 1*) for 10 texture classes are represented (*Fig. 2*).

Table 1 The subsets of dataset

Subset	N	%	Origin
1	111	30.1	undisturbed samples of 9 soil profiles of the r*esearch* program of the *Hungarian* National Long-term Fertilization (Hernádi and Makó, 2011a)
2	123	33.3	undisturbed samples of 12 soil profiles of an investigation for the Hungarian Gas & Oil Company Plc (MOL Rt.) (Makó, 2002; 2005)
3	45	12.2	disturbed samples of mineral mixture series (Hernádi et al., 2011b) (Makó and Marczali, 1999)
4	60	16.3	disturbed samples of aggregate series separated from the upper „A" layer of selected soils (2.0 mm>, 1.0 mm>, 0,5 mm>, 0,25 mm> and 0,056 mm>) (Makó and Elek, 2006)
5*	30	8.1	disturbed samples investigated in the course of the TÁMOP-4.2.1/B-09/1/KONV-2010-0003 Mobility and Environment Project (Hernádi et al., 2011)
Sum	369	100	

* before fluid retention measurements, samples were held 24hr in water for desaggregation and dried on 40°C/24hr

Table 2 Soil properties in the dataset

	N	N%	water retention			NAPL retention		
			vol.%	N	N%	vol.%	N	N%
Surface area (BET) ($cm^2\ g^{-1}$) (Brunaer et al., 1938)	144	39.0	pF 0.0	308	83.5	0 mbar	245	66.4
Plasticity according to Arany (%) (Buzás, 1993)	339	91.9	pF 0.2	170	46.1	2 mbar	361	97.9
Particle size distribution (%)*	312	84.6	pF 1.0	111	30.1	20 mbar	242	65.6
Particle size distribution (%) **	369	100.0	pF 1.3	59	16.0	50 mbar	245	66.4
Organic matter content (%) ***	369	100.0	pF 1.5	279	75.6	100 mbar	116	31.4
$CaCO_3$ (%)***	369	100.0	pF 1.7	59	16.0	150 mbar	245	66.4
Acidity (Na_4OAc) (cmol (+) kg^{-1})***	66	17.9	pF 2.0	156	42.3	200 mbar	116	31.4
Acidity (KCl) (cmol (+) kg^{-1})***	66	17.9	pF 2.2	59	16.0	400 mbar	245	66.4
Salt content (mass %)***	93	25.2	pF 2.3	111	30.1	500 mbar	116	31.4
Exchangeable Na (cmol(+)/kg)***	87	23.6	pF 2.5	168	45.5	1000 mbar	361	97.8
Base saturation (w %)***	24	6.5	pF 2.6	59	16.0	1500 mbar	30	8.1
Aggregate size distribution (%)*	105	28.5	pF 2.7	111	30.1			
pH (distilled water 1 : 2,5)***	273	74.0	pF 3.0	59	16.0			
pH (KCl suspension 1 : 2,5)***	189	51.2	pF 3.2	30	8.1			
Bulk density (water) (g cm^{-3})*	348	94.3	pF 3.4	111	30.1			
Bulk density (NAPL) (g cm^{-3})*	369	100.0	pF 4.2	279	75.6			
			pF 6.2	369	100.0			

* Physical properties of soils were determined according to MSZ 08 0205:1978 standard (PSD – pipette method)
**PSD of soils is determined according to ISO 11277:1995 with pipette method, after the total desaggregation of soil samples (eliminating the soil organic matter and $CaCO_3$ content and the iron oxides).
***Chemical properties of soils were determined according to MSZ 08 0206/2:1978 standard

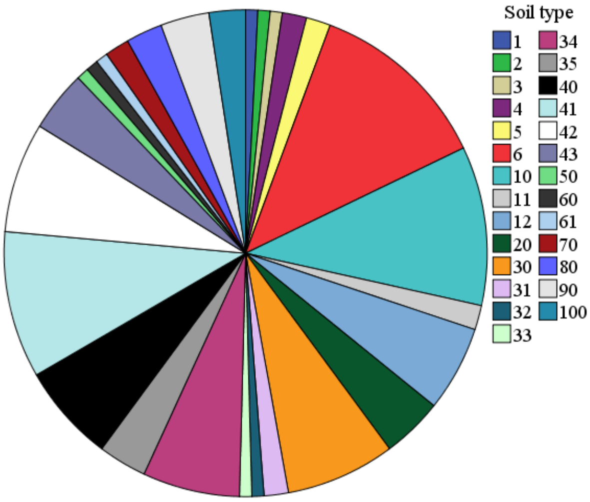

Fig. 1 WRB (World Reference Base) soil types, clay minerals and mineral mixtures in the dataset (1: Benite, 2: Kaolin, 3: Illite, 4: Loess, 5: Pannon sand, 6: Mineral mixtures, 10: Eutric Cambisol, 11: Hortic Terric Cambisol, 12: Dystric Cambisol, 20: Haplic Arenosol, 30: Gleyic Luvisol, 40: Calcic Phaeosem, 41: Luvic Phaeozem, 42: Calcaric Phaeozem, 43: Haplic Phaeozem, 50: Vermic Calcic Chernozem, 60: Vertic Stagnic Solonetz, 61: Orthic Solonetz, 70: Gleyic Vertisol, 80: Calcaric Gleysol, 90: Eutric Regosol, 100: Calcaric Fluvisol)

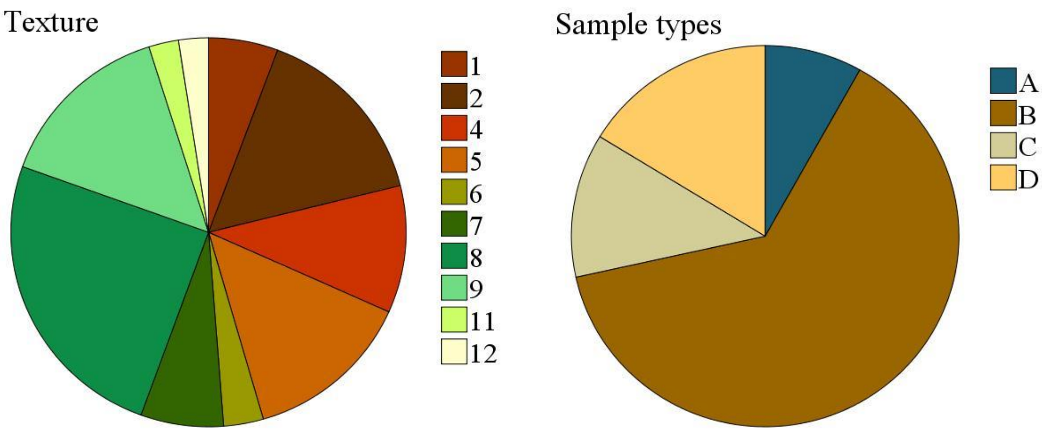

Fig.2 Texture and sample types of soil, clay minerals and mineral mixture samples (1: Clay, 2: Silty clay, 4:Clay loam, 5: Silty clay loam, 6: Sandy clay loam, 7: Loam, 8: Silt loam, 9: Sandy loam, 11:Loamy sand, 12: Sand; A: Disturbed samples, B: Undisturbed samples, C: Mineral mixtures, D: Aggregate series of soil samples)

METHODS

The basic soil properties, used in this investigation were measured according to the Hungarian standards (MSZ08 0205:1978 and MSZ08 0206/2:1978). The texture of soil samples was determined according to the ISSS (International Society of Soil Science) texture triangle (*Table 3*).

The fluid retention measurements were performed with distilled water and a special nonaromatic organic liquid, DUNASOL 180/220 (Hungarian Gas & Oil Company Plc. - MOL Rt.).

Water retention measurements were carried out with porous pressure plate extractors (Soilmoisture Corp. LAB 023). For the purpose of determining the NAPL retention of the samples, a modified version (as described by Makó, 1995) of these pressure plate extractors were used.

Samples showing extremely high or low NAPL retention values were selected with outlier detection. Only the samples, were the casewise diagnostic of the preliminary linear regression between soil parameters and fluid retention values showed significant differences with more than two times the standard deviation at a given pressure level, were eliminated. Then only the records where both the water and NAPL retention were measured were retained. Therefore, the final dataset contained 316 samples.

The fluid retention measurements were performed with distilled water and a special nonaromatic organic liquid, DUNASOL 180/220 (Hungarian Gas & Oil Company Plc. - MOL Rt.).

Water retention measurements were carried out with porous pressure plate extractors (Soilmoisture Corp. LAB 023). For the purpose of determining the NAPL retention of the samples, a modified version (as described by Makó, 1995) of these pressure plate extractors were used.

Table 3 Descriptive statistics of the selected basic soil properties used in this study

	Method	Unit	Mean	SD	SE	Med.	Min.	Max.	N
Clay	MSZ 08 0205:1978	w %	27.58	0.95	16.75	24.08	1.25	67.67	369
Silt			44.22	0.91	16.11	44.39	0.05	78.50	369
Sand			28.21	1.43	25.20	21.43	0.63	98.62	369
OM	MSZ 08 0206/2:1978	w %	1.43	0.06	1.24	1.21	0.00	5.66	369
$CaCO_3$	MSZ 08 0206/2:1978	w %	5.57	0.43	8.26	0.16	0.00	30.71	369
Bulk density (water)*	MSZ 08 0205:1978	g cm^{-3}	1.43	0.01	0.19	1.44	0.89	2.20	369
Bulk density* (NAPL)	MSZ 08 0205:1978	g cm^{-3}	1.42	0.01	0.26	1.45	0.80	1.99	369

*Samples for water and NAPL retention measurements had different bulk density.

Samples showing extremely high or low NAPL retention values were selected with outlier detection. Only the samples, were the casewise diagnostic of the preliminary linear regression between soil parameters and fluid retention values showed significant differences with more than two times the standard deviation at a given pressure level, were eliminated. Then only the records where both the water and NAPL retention were measured were retained. Therefore, the final dataset contained 316 samples.

For fitting the NAPL retention curves the van Genuchten equation with three parameters (Eq. 1) (van Genuchten et al., 1980) were used as initial values in nonlinear regression (SPSS 13.1, Nonlinear regression/quadratic sequential programming) with the average fitting parameters proposed by Carsel and Parish (1988) (used in e.g. RETZ and HSSM models).

$$\theta_{(h)} = \frac{\theta_s}{1+\left((\alpha * h)^n\right)^{1-\frac{1}{n}}} \tag{1}$$

where: $\theta_{(h)}$ is the volumetric fluid content at potential h (kPa); θ_s,is the saturated fluid content; α, and n are fitting parameters.

Afterwards, NAPL retention values for 10 pressure levels (0.01, 0.3, 10, 33, 100, 200, 330, 1000, 15849 and 1584893 mbar) were calculated.

The accuracy of fitting was verified by the calculation of Pearson R^2, RMSE (Eq. 2) and ZAPF values (Eq. 3) (Rajkai, 2004).

$$RMSE = \sqrt{\frac{\sum_{1}^{N}(y_i - \hat{y}_i)^2}{N}} \tag{2}$$

where: y_i *is the fitted NAPL retention,* $\hat{y}_i$ *the estimated NAPL retention and* N *is the number of samples.*

In case of ZAPF values (Eq. 3.), the lower |ZAPF| than measurement error (ME) of NAPL retention was regarded as sufficiently fitted (as suggested by Rajkai, 2004).

$$ZAPF = \frac{\sum_{i=1}^{n}\left|\theta_e - \theta_m\right|}{n} \tag{3}$$

where: θ_m is the measured fluid retention value θ_e estimated fluid retention values; n is the number of fitting point of fluid retention curves.

Before creating PTFs the dataset was randomly split with the ratio of 90:10 (calibration and validation data set), thus we had opportunity to investigate the accuracy and the also the reliability of estimations (*Table 4*). In further investigations the fitted fluid retention data was used, which enabled us to compare the retention data determined with both fluid types measured at different pressure levels. Relationships between basic soil properties and fluid retentions, as well as between soil properties and fitting parameters were preliminarily investigated with principal component analysis (PCA) (SPSS 13.1/Varimax rotation with Kaiser normalization).

Subsequently, parametric PTFs were built as class PTFs for selected four texture groups (silty clay, silty clay loam, silt loam and sandy loam) with multiple linear regression (SPSS 13.1 Linear regression/Stepwise method). In linear regression, bulk density, clay, silt, $CaCO_3$ and organic matter content and their derived values, such as the logarithmic, reciprocal, squared values and their interactions were used as independent variables (as suggested by Wösten, 1995).

The NAPL retention of the soil samples were predicted with a modified fitting method proposed by Lenhard and Parker (1987) (Eq. 4) which was a combination of the Leverett equation (1941) and a four-parametric fitting procedure suggested by van Genuchten et al. (1980).

$$S = \left[\frac{1}{1+(\beta * \alpha * P)^n}\right]^m \tag{4}$$

where: S is the fluid saturation; P is the pressure head; β is a scaling factor calculated from the interfacial tension of the phase pair and α, n and m are fitting parameters.

In scaling method the m = 1-1/n constraints was supposed as suggested by van Genuchten (1991).

In order to compare the accuracy and reliability of the different predictions, the RMSE and R^2 values were determined. The NAPL saturation values of soil samples were also calculated on the validation test in order to make a preliminary investigation to compare the reliability of PTFs with scaling method in predicting NAPL saturation values, versus the pressure.

Table 4 Statistics of the fitted fluid retention values in calibration and validation datasets

Pressure (mbar)	Calibration set						Validation set					
	N = 264						N = 26					
	Mean	SE	SD	Med.	Min.	Max.	Mean	SE	SD	Med.	Min.	Max.
NAPL retention												
0.0	41.0	1.0	10.0	41.0	22.0	68.0	38.0	2.0	10.0	37.0	25.0	57.0
0.33	41.0	1.0	10.0	41.0	22.0	68.0	38.0	2.0	10.0	36.0	25.0	57.0
10	36.0	1.0	10.0	35.0	17.0	64.0	33.0	2.0	8.0	32.0	21.0	49.0
33	31.0	1.0	8.0	31.0	9.0	62.0	28.0	1.0	5.0	29.0	20.0	38.0
100	25.0	0.0	6.0	25.0	5.0	60.0	24.0	1.0	5.0	24.0	16.0	33.0
200	22.0	0.0	6.0	21.0	3.0	55.0	22.0	1.0	6.0	22.0	10.0	32.0
330	20.0	0.0	6.0	19.0	2.0	49.0	21.0	1.0	6.0	22.0	7.0	32.0
1000	17.0	0.0	6.0	16.0	1.0	34.0	18.0	1.0	7.0	18.0	3.0	30.0
15849	11.0	0.0	7.0	11.0	0.0	29.0	14.0	1.0	7.0	14.0	0.0	26.0
1584893	7.0	0.0	6.0	6.0	0.0	22.0	10.0	1.0	6.0	9.0	0.0	21.0
Water retention												
0.0	50.0	1.0	11.0	47.0	27.0	81.0	49.0	2.0	9.0	47.0	37.0	70.0
0.33	50.0	1.0	11.0	47.0	26.0	81.0	49.0	2.0	9.0	47.0	37.0	70.0
10	48.0	1.0	11.0	46.0	16.0	78.0	47.0	2.0	9.0	46.0	36.0	69.0
33	45.0	1.0	11.0	43.0	12.0	73.0	45.0	2.0	10.0	42.0	31.0	67.0
100	41.0	1.0	12.0	39.0	5.0	73.0	42.0	2.0	10.0	40.0	26.0	63.0
200	37.0	1.0	12.0	36.0	1.0	72.0	39.0	2.0	10.0	39.0	23.0	62.0
330	35.0	1.0	12.0	33.0	0.0	70.0	37.0	2.0	10.0	38.0	21.0	61.0
1000	29.0	1.0	11.0	27.0	0.0	65.0	32.0	2.0	10.0	32.0	17.0	56.0
15849	17.0	0.0	8.0	16.0	0.0	41.0	19.0	1.0	7.0	19.0	7.0	37.0
1584893	7.0	0.0	4.0	6.0	0.0	19.0	8.0	1.0	3.0	8.0	1.0	15.0

RESULTS AND DISCUSSION

Fitting van Genuchten equation

As the determination coefficient showed NAPL retention curves can be accurately fitted with the hydraulic function proposed by van Genuchten (1980). Moreover, according to the ZAPF values, the differences between the fitted and measured NAPL retention values were lower than the measurement error in more than 98.4% of the samples (*Table 5*).

Table 5 Statistics of the fitted van Genucthen equation to the measured NAPL retention values

Equation	ZAPF (<0,75 – ME*)		RMSE	R	R^2
	N	Frequency (%)			
van Genuchten (3 parameter)	2286.0	98.4	1.2	0.99	0.99

* ME is the average error of NAPL retention measurement

Principal component analysis

Principal components were extracted from 17 variables which showed strong correlations between the selected soil properties. The maximum possible variance of 90.92% can be explained by the three components. Scores of component I (CI) give information about 69.56% scores of component II (CII) about 15.48% and scores of component III (CIII) about 5.87% of the variance of variables.

After Kaiser normalization, all components with eigenvalues under 1.0 were eliminated, and only three components were taken into account. In *Table 6* and *7*, component scores less than the variance explained criteria (0.4) are shown in gray. Communality values show, that all the variances of each correlated variables might be accounted for by the components. The finer textured fractions belonged to the same component with the NAPL retention at higher pressure levels, where the organic liquid can be retained against gravity mainly by capillary and adsorptive forces. The joint variation of these soil properties were affected by CI as the unobserved latent variable. Moreover, the component scores of finer particle size fractions exceeded the 0.4 criteria only in CI.

Differences in the role of components on the NAPL retention of soils were observed. Increasing value of component scores with increasing pressure can refer to the strong correlations between particle size distribution and fluid retention, especially in the higher pressure range.

Higher component scores of CII were observed than CI on the 0-50mbar pressure range. In CII only the fluid retention less than 150 mbar were represented while the component scores of CI in case of NAPL retention at higher than 400 mbar exceed the 0.9 value (*Table 6.)* This difference might be resulted from reaching a threshold pressure range, which can be in the range of 20-50 mbar. At this pressure range the larger pores might be drained by gravity during NAPL retention measurements with Dunasol 180/220, as similar results were observed by Makó et al. (2011a). This is lower than the pressure of the field capacity of soils, which is referred approximately 400 mbar (pF 2.5) in the case of water retention.

This is might not be contradicted to the results of other researchers for water retention (Vereecken et al., 1989; Saxton and Rawls, 2006). Vereecken et al., (1989) resulted that at pressure higher than 1500 kPa water retention is determined mainly by texture then in lower pressure range the fluid retention strongly influenced by the aggregation of soil particles and organic matter content as well.

However, in case of NAPL retention the variance of organic matter content affects the NAPL retention but in decreasing rate with increasing pressure in CI and CII, and calcium carbonate content of samples appeared only in CIII.

Variance in bulk densities can be explained by all three components, thus it may influence the NAPL retention at every pressure level. In all cases bulk density had negative correlation with other variables which refers inverse relationships between bulk density and fluid content of soils samples.

More complex and less correlation between parameters of water and NAPL retention curves related to basic soil properties were observed. In case of water retention only the 55.59% of the variance could be explained by the components as compared to NAPL retention (76.55% (*Table 7*).

Parameters α, n and θ_s have substantial roles in CI of NAPL retention in contrast to the parameters of water retention curves which belong to separated principal components. Descriptive statistics show (*Table 4*), that the mean saturated water content is significantly higher (50 vol %) than the possible average maximum NAPL retention (41 vol %), which might influence the variability of the other two fitting parameters.

Parameters α and n are shape factors, strongly influenced by the pore size distribution of soils and structure (van Genuchten et al., 1991; Vereecken et al., 1989). However, there is a difference in effective pore size distribution of soils saturated by fluids with various chemical and physical properties. In addition, interactions between water and the solid phase during drainage and imbibition processes, such as swelling and shrinking, have not occurred in case of the soil pores filled with NAPLs. Thus, the soil properties which might influence the shape of fluid retention curves might be difference.

Strong correlation between the fitting parameters was experienced when the van Genuchten equation was fitted to NAPL retention data. The correlation between α and θ_s of the selected four texture group varied between 0.553-0.887. This might refers the inability of using the same initial parameters or boundary conditions for fitting NAPL and water retention curves.

Lower communality values were experienced in the n and θ_s values of water retention curves than those of NAPL retention curves. This could be resulted from the variability in NAPL retention of soils might be explained better with the selected basic soil properties (communalities of n was 0.821, and θ_s was 0.739) than in case of water retention (where the cumulated variance was only 0.282 and 0.298).

Table 6 Components of variance and scores of the principal component analysis

	Variables	CI	CII	CIII	Communalities
Basic soil properties	0.0002 mm >	0.93	0.22	-0.19	0.94
	0.005-0.002 mm	0.87	0.36	0.11	0.89
	0.01-0.005 mm	0.91	0.30	0.18	0.94
	0.02-0.01 mm	0.75	0.35	0.50	0.93
	0.05-0.02 mm	0.02	0.41	0.88	0.95
	0.05-0.25 mm	-0.94	0.05	-0.33	0.99
	Organic matter (w %)	0.56	0.46	0.02	0.53
	$CaCO_3$ (w %)	-0.07	0.05	0.95	0.91
	Bulk density (g cm^{-3})	-0.51	-0.53	-0.53	0.82
NAPL retention	0mbar (w %)	0.48	0.66	0.46	0.88
	2mbar (w%)	0.49	0.65	0.45	0.87
	20mbar (w %)	0.46	0.76	0.39	0.94
	50mbar (w %)	0.56	0.59	0.54	0.96
	150mbar (w %)	0.78	0.46	0.40	0.98
	400mbar (w %)	0.91	0.34	0.14	0.96
	1000mbar (w %)	0.93	0.34	0.04	0.98
	1500mbar (w %)	0.90	0.41	-0.05	0.98
Statistics	Variance %	69.56	15.48	5.87	

Table 7 Components of variance and scores of the principal component analysis with parameters of fluid retention curves

Variables	Water retention				NAPL retention			
4	CI	CII	CIII	Communalities	CI	CII	CIII	Communalities
a	0.025	-0.072	0.753	0.665	-0.738	0.451	-0.041	0.749
n	-0.155	0.581	-0.101	0.282	0.468	-0.775	0.021	0.821
θ_s	-0.602	0.171	-0.228	0.298	0.841	0.155	0.085	0.739
Clay content (w %)	0.847	-0.163	-0.237	0.933	0.409	0.657	-0.455	0.806
Silt content (w %)	0.469	0.478	0.077	0.932	0.097	0.815	0.337	0.788
Organic matter (w %)	0.014	0.123	0.784	0.579	-0.187	0.074	0.643	0.454
$CaCO_3$ (w %)	0.084	0.801	0.137	0.722	0.426	0.03	0.726	0.709
Bulk density (g cm^{-3})	-0.638	-0.315	-0.034	0.686	-0.889	0.016	0.136	0.808
Variance %	22.768	17.792	22.768		33.191	29.179	14.184	
Cumulated variance %	22.768	40.560	55.599		33.191	62.370	76.554	
Bartlett's Test of Sphericity			0.000		0.000			
Kaiser-Meyer-Olkin Measure of Sampling Adequacy			0.597		0.634			

In both cases bulk density is presented in CI with negative effect on saturated fluid content, which can be a consequence the bulk density and saturated fluid content are closely related.

In case of water retention the role of organic matter content is apparent only in CIII separately from clay content. This can be a consequence of the effect of high clay content might mask the effects of increasing organic matter for water retention, as Saxton and Rawls (2006) and Rawls et al. (2003) experienced as well.

Similar relationships observed between organic matter content and the fitting parameters of water retention curves than Petersen et al. (1968) and Vereeckeen et al. (1989) that is the organic matter content of soils affects primary the α parameter of water retention curves, which refers the fluid retention values at inflection point. In contrary with texture, organic matter content of soils influence the position of the retention curves, rather than its shape.

Communality values showed the same important role of calcium carbonate (0.722, 0.709) and organic matter (0.579, 0.454) on water and NAPL retention. However, a separated component (CIII) was composed by organic matter, $CaCO_3$ and clay content in case of NAPL retention data. In addition the role of clay content on water retention seemed to be higher than its role on NAPL retention. All of these observed relationships refer to the different rate of interactions between various fluids and the solid phase considering these express mainly the linear relationships between these variables.

In spite of the cumulated variance was only 55.59 % in case of water retention, the Kaiser-Meyer-Olkin test showed the applicability of PCA, sampling adequacy (as it refers that the correlations between pairs of variables can be explained with other variables) was higher than 0.5, and the Barthlett's test indicated significant correlation between variables ($p<0.05$).

Pedotransfer functions

Accurate prediction was given by classPTFs for the selected four texture groups for estimating NAPL retention of soils (*Table 8*).

Table 8 Accuracy of PTFs to estimate fitting parameters of NAPL retention curves

Parameter	R^2	RMSE
α	0.978	13.59
n	0.900	17.75
θ_s	0.988	7.50

According to the RMSE values classPTFs give more accurate estimations to saturated fluid content than the shape parameters α or n. The lower efficiency of estimating parameter α might be caused by the strong correlations experienced between parameters α and θ_s and might also resulted from the weaker correlation with independent variables.

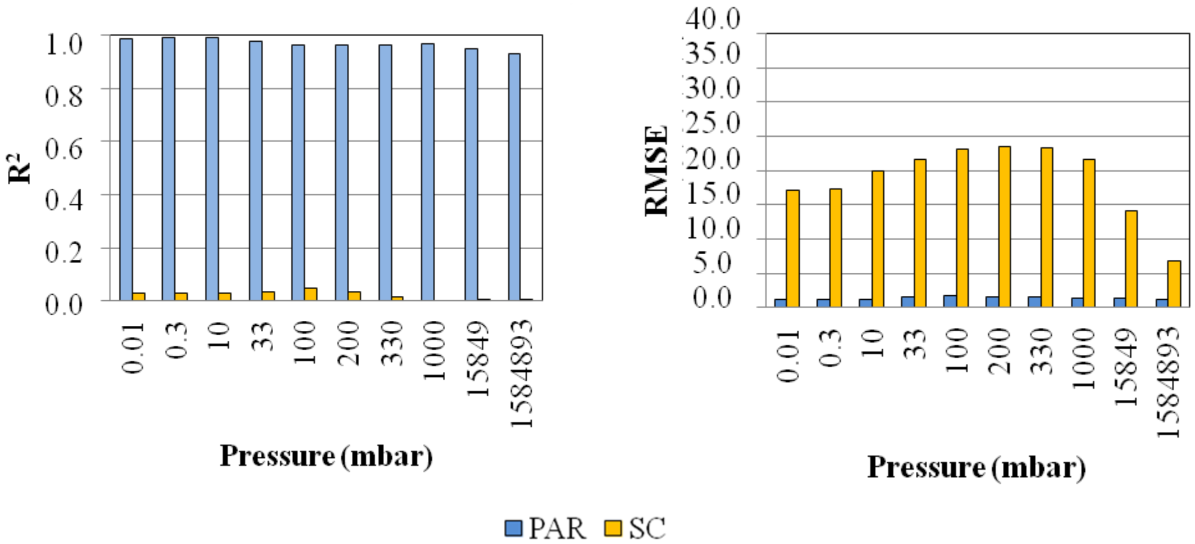

Fig. 3 Accuracy of estimating volumetric NAPL retention values with different estimation methods
PAR: the parametric method based on basic soil properties and their derived values;
SC: Scaling method (Lenhard and Parker, 1987)

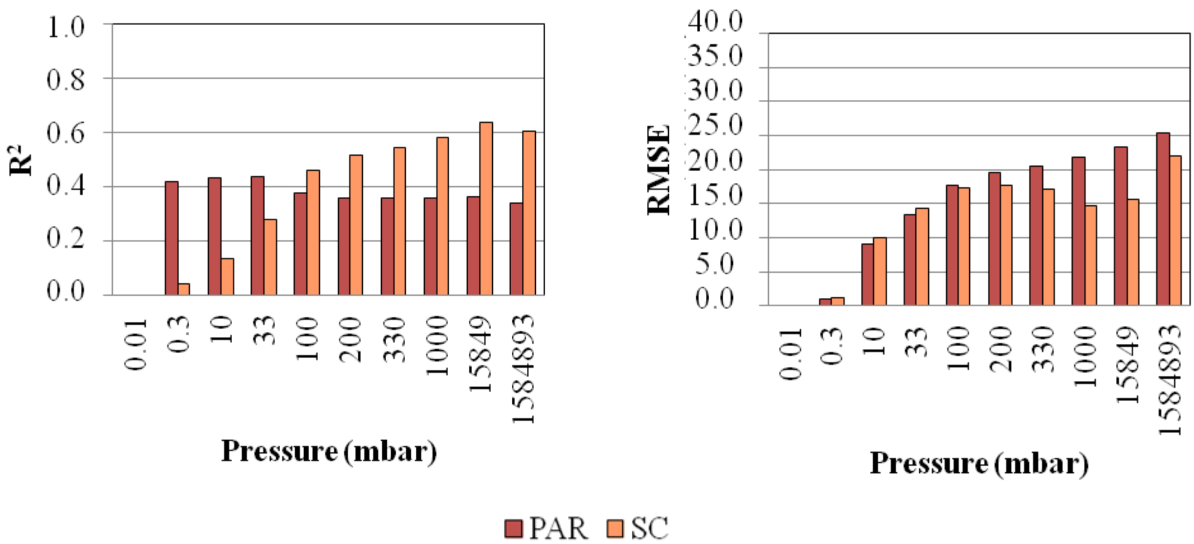

Fig. 4 Reliability of estimating NAPL saturation values with PTF and scaling methods
PAR: the parametric method based on basic soil properties and their derived values;
SC: Scaling method (Lenhard and Parker, 1987)

The applicability of PTFs to predict the NAPL retention values was proven by the statistical attributes (R^2 and RMSE). In comparison the accuracy of parametric PTFs with the scaling method, NAPL retention might be better predicted with PTFs at a given pressure level (*Fig. 3*).

Applicability of these estimation methods for predicting NAPL saturation values of soils was compared with the determination of R^2 and RMSE values on validation data (10 % of the dataset, 26 samples). Lower reliability of PTFs was found to estimate the NAPL saturation with PTFs than the scaling method at higher than 100 mbar pressure level. However, PTFs could provide estimates sufficiently reliable for NAPL retention at all pressure level, especially near saturation (*Fig. 4*). On *Fig. 4* the statistics of saturated fluid content were not presented because their values are 1 for both fluids.

Supposedly, scaling method gave increasing better estimation for NAPL retention as a consequence of decreasing difference between fluid retentions with NAPL and water. As it shown formerly from descriptive statistics (*Table 4*), the difference between the average water and NAPL retention of the investigated soil samples was 26 vol. % at 100 mbar but practically 0 vol. % at 1584893 mbar.

At the same time RMSE values indicated lower difference in reliability of estimation methods for predicting the NAPL saturation.

Only moderate reliability of PTFs was found when predicting the volumetric NAPL retention values (*Fig. 5*). In spite of the fact that higher RMSE values were observed for PTFs in lower pressure ranges, the R^2 values show that the reliability of PTFs exceed that of the scaling method at lower pressure levels (< 330 mbar) and is al-

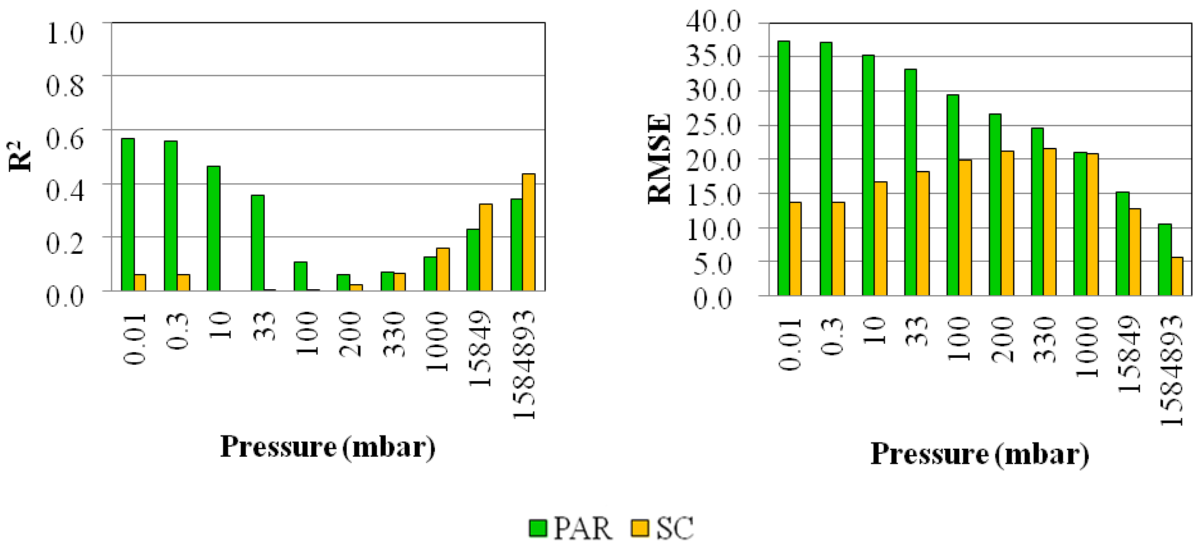

Fig. 5 Reliability of estimating volumetric NAPL retention values with PTF and scaling methods
PAR: the parametric method based on basic soil properties and their derived values;
SC: Scaling method (Lenhard and Parker, 1987)

most equal to it at pressure higher than 1000 mbar (*Fig. 5*). Moreover both the decreasing R^2 values of the scaling method and the increasing RMSE values in the mid-pressure range (near the inflection point of the NAPL retention curves) suggest the inability of scaling methods for predicting NAPL retention of soil based on water retention.

CONCLUSION

In the last fifteen years several investigations were performed in order to create estimation methods to predict the NAPL retention of soils. However, the results of these measurements have not been collected into comprehensive and well established databases, yet.

Building PTFs using the data of easily measured soil property data might be a promising alternative for predicting the NAPL retention of soils. By the creation of an organic liquid retention database representative for most soil types we can generate well established point and parametric PTFs based on basic soil properties. After the validation of PTFs with the results of column experiments it might be inserted to transport models. Moreover, databases of soil hydraulic and basic properties which have already been used in developing PTFs for estimating water retention might be applicable to create NAPL pollution sensitivity maps.

In our study an organic liquid retention dataset from recently measured NAPL and water retention data of soil and mineral mixture samples was created. The fluid retention measurements were carried out with the pressure plate extractor method. Furthermore, the basic soil properties, such as particle size distribution, organic matter and $CaCO_3$ contents, and bulk density, etc. were also measured.

According to our results, the van Genuchten equation, having three parameters, could be perfectly fitted to the measured NAPL retention data as well as to water retention curves.

Preliminary analysis of the filtered dataset of 316 soil samples with PCA showed strong correlations between the selected basic soil properties as well as fluid retentions of water and NAPL. The important role of organic matter and $CaCO_3$ content in fluid retention of soils beyond the particle size distribution and bulk density was proved.

The estimation methods for predicting the volumetric NAPL retention values of soils PTFs gave better results than the scaling method. As the parameters of NAPL retention curves were correlated (e. g. fitting with nonlinear regression might offer more than one possible solution), further research is needed to investigate the role of initial parameter values in fitting hydraulic functions to NAPL retention data.

Reliability of PTFs and the scaling method to predict the saturation values versus the pressure were compared. Similar reliable estimations found when predicting the NAPL saturation values with scaling method and with PTFs. However, PTFs has been successfully applied to estimate the volumetric NAPL retention, more reliable than the scaling method.

Acknowledgements

The authors gratefully acknowledge Ágota Horel and Gábor Barton, for their assistance in preparing the manuscript. We would also like to thank Jenőné Borbély, for the support in laboratory experiments at Pannon University.

This research was supported by the European Union and the State of Hungary, co-financed by the European Social Fund in the framework of TÁMOP-4.2.4.A/ 2-11/1-2012-0001 'National Excellence Program'.

References

Acutis, M., Donatelli, M. 2003. SOILPAR 2.00: Software to estimate soil hydrological parameters and functions. *European Journal of Agronomy* 8 (3–4), 373–377. DOI: 10.1016/s1161-0301(02)00128-4

Beckett, G.D., Joy, S. 2003. Light Non-Aqueous Phase Liquid (LNAPL) Parameters Database. Version 2.0. Users Guide. American Petroleum Institute. Publ. 4731. Washington, DC.

Brooks, R. H., Corey, A. T., 1964. Hydraulic properties of porous media. Hydrology Paper No. 3. Colorado State University. Fort Collins, Colorado.

Brunauer, S., Emmett, P. H., Teller, E. 1938. Adsorption of gases in multimolecular layers. *Journal of the American Chemical Society 60 (2) 309–319.* DOI: 10.1021/ja01269a023

Brutsaert, W. 1966. Probably laws for pore size distributions. *Soil Science* 101 85–92. DOI: 10.1097/00010694-196602000-00002

Buzás, I. 1993. Manual for soil and agrochemical analysis. INDA Kiadó, Budapest. (in Hungarian)

Carsel, R.F., Parrish, R.S. 1988. Developing joint probability distribution of soil water retention characteristics. *Water Resources Research* 24 755–769. DOI: 10.1029/wr024i005p00755

Chen, J., Hopmanns, J. W., Grismer, M. E. 1999. Parameter estimation of two-fluid capillary pressure-saturation and permeability functions. *Advances in Water Resources* 22, 479–493. DOI: 10.1016/s0309-1708(98)00025-6

DePastrovitch, T. L., Baradat, Y., Barthel, Y., Chiarelli, A., Fussel, D. R., 1979. Protection of groundwater from oil pollution. CONCAWE Rep. No 3/79. CONCAWE (Conservation of Clean Air and Water – Europe). The Hague. The Netherlands.

Fodor, N., Rajkai, K. 2011. Computer program (SOILarium 1.0) for estimating the physical and hydrophysical properties of soils from other soil characteristics. *Agrokémia és Talajtan* 60, 27–40.

Hernádi, H., Makó, A. 2010. Predicting oil retention of soils polluted with hydrocarbon derivates with pedotransfer functions. Mérnökgeológia-kőzetmechanika. Műegyetem Kiadó. Budapest. (in Hungarian)

Hernádi, H., Makó, A. 2011a. Predicting the oil retention of soils with different methods. In Farsang, A., Ladányi, Zs. (eds.) *Talajvédelem* supplementum (3-4), 363–371 (In Hungarian)

Hernádi, H., Makó, A., Kovács, J., Csatári, T. 2011b. Nonaqueous-phase liquid retention of mineral mixture series containing different clay minerals. *Communications in Soil Science and Plant Analysis* 44, 390–396. DOI: 10.1080/00103624.2013.742314

Lamorsky, K., Pachepsky, Y., Sławińsky, C., Walczak, R.T. 2008. Using support vector machines to develop pedotransfer functions for water retention of soils in Poland. *Soil Science Society America Journal* 72, 1243–1247. DOI: 10.2136/sssaj2007.0280n

Lenhard, R .J., Parker, J. C., 1987. Measurement and prediction of saturation-pressure relations in three phase porous media systems. *Journal of Contaminant Hydrology* 1. 407–424. DOI: 10.1016/0169-7722(87)90017-9

Leverett, M. C. 1941. Capillary behavior in porous solids. *Transactions of the Society of Petroleum Engineers*. *American Institute of Mechanical Engineers* 142, 152–169.

Lilly, A. 2010. A hydrological classification of UK soils based on soil morphological data. 19th World Congress of Soil Science, Soil Solutions for a Changing World 1 – 6 August 2010, Brisbane, Australia. Published on DVD.

Makó, A. 1995. Interactions between the porous phase of soils and organic liquids. PhD dissertation. Keszthely, Hungary. (in Hungarian)

Makó, A. 2002. Measuring and estimating the pressure-saturation curves on undisturbed soil samples using water and NAPL. *Agrokémia és Talajtan* 51, 27–36.

Makó, A. 2005. Measuring the two-phase capillary pressure-saturation curves of soil samples saturated with nonpolar liquids. *Communication in Soil Science and Plant Analysis* 36, 439–453. DOI: 10.1081/css-200043170

Makó A., Elek, B. 2006. Comparison of soil extraction isotherms of soil samples saturated with nonpolar liquids. *Water, Air and Soil Pollution* 6, 331–342. DOI: 10.1007/s11267-005-9026-x

Makó, A., Hernádi H. 2013. Hydrocarbon derivates in soils: Soil physical researches. Monography. (in Hungarian)

Makó, A., Marczali, Zs. 1999b. Laboratory measurement of the soils fluid retention concerning the organic liquid retention. XIII. Országos Környezetvédelmi Konferencia és Szakkiállítás. Siófok. 14-16. szept. 1999. 147-153. (In Hungarian)

Makó, A., Máté, F., Németh, T., Hernádi H. 2004. The temperature-dependence of the NAPL retention of different soils. 12th International Poster Day and Institute of Hydrology Open Day: Transport of water, chemicals and energy in the soil-plant-atmosphere system. 25th November 2004. Bratislava. Slovak Republik.

Makó, A., Rajkai, K., Tóth, G., Hermann, T. 2005. Estimating soil water retention characteristics from the soil taxonomic classification and mapping informations: consideration of humus categories. *Cereal Research Communications* 34, 199–201. DOI: 10.1556/crc.33.2005.1.27

McBratney, A. B., Minasny B., Cattle S. R., Vervoort R. W. 2002. From pedotransfer functions to soil inference systems. *Geoderma* 109, 41–73. DOI: 10.1016/s0016-7061(02)00139-8

Minasny, B., McBratney, A. B., Bristow, K. L. 1999. Comparison of different approaches to the development of pedotransfer functions for water-retention curves. *Geoderma* 93, 225–253. DOI: 10.1016/s0016-7061(99)00061-0

Minasny, B., J., Hopmans, J. W., Harter, T., Eching, S. O., Tuli, A., Denton, M. A. 2004. Neural networks prediction of soil hydraulic functions for alluvial soils using multistep outflow data. *Soil Science Society of America Journal* 68, 417–429. DOI: 10.2136/sssaj2004.4170

Nemes, A., Schaap, M. G., Wösten, J. H. M. 2003. Functional evaluation of pedotransfer functions derived from different scales of data collection. *Soil Science Society of America Journal* 67, 1093–1102. DOI: 10.2136/sssaj2003.1093

Nemes, A., Roberts, R. T., Rawls, W. J., Pachepsky, Y. A., van Genuchten, M. T. 2008. Software to estimate -33 and -1500 kPa soil water retention using the non-parametric k-nearest neighbor technique. *Environmental Modelling & Software* 23 (2), 254–255. DOI: 10.1016/j.envsoft.2007.05.018

Pachepsky, Y. A., Rawls, W. J. (Eds.) 2004. Development of Pedotransfer Functions in Soil Hydrology. Developments in Soil Science. Elsevier. Amsterdam.

Parker, J.C., Lenhard, R.J., Kuppusamy,T. 1987. A parametric model for constitutive properties governing multi-phase flow in porous media. *Water Resources Research* 23, 618–624. DOI: 10.1029/wr023i004p00618

Petersen, G. W., Cunnungham, R. L., Matelski, R.P. 1968. Moisture characteristics of Pennsylvania soils II. Soil factors affecting moisture retention within a textural class - silt loam. *Soil Science Society America Proceedings* 32, 866–870. DOI: 10.2136/sssaj1968.03615995003200060042x

Rajkai, K. 1988. Relationships between water retention and different soil properties. *Agrokémia és Talajtan* 36–37, 15–30. (In Hungarian)

Rajkai K. 2004. The quantity, distribution and transport of water in soil. RISSAC, Research Institute for Soil Sciences and Agricultural Chemistry. Budapest. (In Hungarian)

Rajkai, K., Kabos, S., van Genuchten, M. T., Jansson, P. E. 1996. Estimation of water-retention characteristics from the bulk density and particle-size distribution of Swedish soils. *Soil Science* 161, 832–845. DOI: 10.1097/00010694-199612000-00003

Rajkai, K., Kabos, S., van Genuchten, M. T., 2004. Estimating the water retention curve from soil properties: Comparison of linear, nonlinear and concomitant variable methods. Soil and Tillage Research 79. (2) 145–152. DOI: 10.1016/j.still.2004.07.003

Rathfelder, K., Abriola, L. M. 1996. The influence of capillarity in numerical modelling of organic liquid redistribution in two-phase systems. *Advances in Water Resources* 21 (2), 159–170. DOI: 10.1016/s0309-1708(96)00039-5

Rawls, W. J., Brakensiek, D. L., 1985. Prediction of soil water properties for hydrologic modeling. In: Watershed Management in the 1980s. Proceeding of Symposium of Irrig. Drainage Div., Denver, CO., April 30–May 1, 1985. American Society pf Civil Engineers. New York. 293-299.

Rawls, W. J., Pachepsky, Y. A., Shen, M. H., 2001. Testing soil water retention estimation with the MUUF pedotransfer model using data from the southern United States. *Journal of Hydrology* 251. 177–185. DOI: 10.1016/s0022-1694(01)00467-x

Rawls, W.J., Pachepsky, Y. A., Ritchie, J. C., Sobecki, T. M., Bloodworth, H. 2003. Effect of Soil Organic Carbon on Soil Water Retention. Geoderma 1 (16), 61–71. DOI: 10.1016/s0016-7061(03)00094-6

Saxton, K. E., Rawls, W. J. 2006. Soil Water Characteristic Estimates by Texture and Organic Matter for Hydrologic Solutions. *Soil Science Society America Journal* 70, 1569–1578. DOI: 10.1016/s0016-7061(03)00094-6

Schaap, M. G., Leij, F. J., van Genuchten, M. T. 1999. A bootstrap-neural network approach to predict soil hydraulic parameters. In:

Proceedings of International Workshop Characterization and Measurements of the Hydraulic Properties of Unsaturated Porous Media 1237–1250.

Schaap, M. G., Leij, F. J., van Genuchten, M. T. 2001. Rosetta: A computer program for estimating soil hydraulic parameters with hierarchical pedotransfer functions. *Journal of Hydrology* 251, 163–176. DOI: 10.1016/s0022-1694(01)00466-8

Tietje, O., Tapkenhinrichs, M. 1993. Evaluation of pedo-transfer functions. *Soil Science Society of America Journal* 57, 1088–1095. DOI: 10.2136/sssaj1993.03615995005700040035x

Tóth, B. 2011. Calculation and characterization of water retention of major Hungarian soil types using soil survey information. PhD dissertation. Keszthely. (in Hungarian)

Tóth, B., Makó, A., Rajkai, K., Kele, Sz. G., Hermann, T., Marth, P. 2006. Use of soil water retention capacity and hydraulic conductivity estimation in the preparation of soil water management maps. *Agrokémia és Talajtan* 55, 49–58. DOI: 10.1556/agrokem.55.2006.1.6

Tóth, B., Makó, A., Tóth, G., Farkas, C., Rajkai, K. 2013. Comparison of pedotransfer functions to estimate the van Genuchten parameters from soil survey information. Agrokémia és Talajtan 62 (1), 5-22. (in Hungarian)

van Genuchten M.T. 1980. A closed form equation for predicting the hydraulic conductivity of unsaturated soils. *Soil Science Society American Journal* 44, 892–989. DOI: 10.2136/sssaj1980.03615995004400050002x

Vereecken, H., Maes, J., Feyen, J., Darius, P. 1989. Estimaing the soil moisture retention characteristic from texture, bulk density and carbon content. *Soil Science* 148, 389–403. DOI: 10.1097/00010694-198912000-00001

Wösten, J. H. M., Schuren, C. H. J. E. Bouma, J., Stein, A. 1990. Functional sensivity analysis of four methods to generate soil hydraulic functions. *Soil Science Society of America Journal* 54, 832–836. DOI: 10.2136/sssaj1990.03615995005400030036x

Wösten, J.H.M., Finke, P.A., Janses, M.J.W. 1995. Comparison of class and continuous pedotransfer functions to generate soil hydraulic characteristics. *Geoderma* 66, 227-237. DOI: 10.1016/0016-7061(94)00079-p

Wösten, J. H. M., Lilly, A., Nemes, A., Le Bas, C. 1999. Development and use of a database of hydraulic properties of European soils. *Geoderma* 90, 169–185. DOI: 10.1016/s0016-7061(98)00132-3

Weawer, J. L., Charbeneau, R. J., Tauxe, J. D., Lien, B. K., Provost, J. B. 1994. The hydrocarbon spill screening model (HSSM). 1. US EPA. EPA/600/R-94/039a.

JOURNAL OF ENVIRONMENTAL GEOGRAPHY
Journal of Environmental Geography 7 (1–2), 23–28.
DOI: 10.2478/jengeo-2014-0003
ISSN: 2060-467X

SEDIMENT DYNAMICS IN A SMALL, 2ND ORDER URBAN RIVER AWBA CATCHMENT, IBADAN, NIGERIA

Aderemi Adediji[1*], Olutoyin A. Fashae[2]

[1]Department of Geography, Obafemi Awolowo University, Faculty of Social Sciences, 20005 Ile-Ife, Nigeria
[2]Department of Geography, University of Ibadan, Faculty of Social Sciences, 200284 Ibadan, Nigeria
*Corresponding author, e-mail: remiadediji2003@yahoo.co.uk

Research article, received 02 February 2014, accepted 06 March 2014

Abstract

The sediment dynamics in a small 2nd order catchment of River Awba in the territory of the University of Ibadan, Nigeria was investigated between January and December 2012. The river was gauged by daily measurements of water level as well as sampling of water for determination of suspended sediment load. In this regard, apart from weekly sample, twelve (12) storm flow events which occurred during the day were sampled for determination of suspended sediment concentration. The results showed that during the storms the suspended sediment concentration varied between 636 mg/l in May and 3641.5 mg/l in September, with a mean of 2136.8 mg/l. Also, the value of monthly suspended sediment yield ranged from 10.85 kg in January to 288.4 kg in October with a mean of 89.5 kg. The variability in monthly sediment load closely followed the trend of monthly rainfall in the study area. However, in order to minimize the storm runoff and sediment load generated from the rainstorms events, the paved surfaces within the study catchment should be grassed with the planting of some few tree species. This could further reduce the rate of floods occurrence.

Keywords: gauged, suspended sediment, rainfall, River Awba catchment

INTRODUCTION

Sediment yield is an indication of soil erosion in the river catchments. It comprises of both suspended and dissolved load discharge through the basin outlet. These materials (both suspended and solute load) are generated from both interfluvial areas and channel subsystems within the drainage basin. For instance, Mwamba and Torres (2009) in United States, stated that rivers are the major source of ocean sediment, and the coastal and marine environment is the ultimate sink for most fluvial systems. Thus, it might seem likely that significant changes in erosion or sediment transport within the drainage basin would be reflected in the changes in sediment delivery in coastal and marine environment. However, the link between sediment dynamics within a fluvial system and sediment load at the river outlet is not always strong or direct. This is because storage opportunities or time lags may make some drainage basins/catchments, slowly responsive or relatively unresponsive to environmental change. However, the amount of material or sediment generated and transported from the basin depends on both physiographic and catchment characteristics such as climate (rainfall), vegetation/land use condition, slope factor, basin area, relief ratio, drainage density and sinuosity index.

In this regard a lot of published studies exist on factors influencing catchment sediment yield in both tropical and extra tropical regions (e.g. Smets et al., 2008; Peng, 2008 in United States; Bracken and Kirkby, 2005; Nadal and Reques, 2008; Nadal et al., 2008; Noenu et al., 2010; Fang et al., 2012; Nu-Fang et al., 2012; Taguas et al., 2013 in Spain; Lopez et al., 2010 in Italy; Wei et al., 2007; Zeng et al., 2008 in China; Adriana et al., 2012 in France; Ogunkoya, 1980; Ogunkoya and Jeje, 1987; Oluwatimilehin, 1991; Jeje et al., 1999; Adediji and Jeje, 2004; Adediji et al., 2013 in Southwestern Nigeria). For example, Slattery and Phillips (2010) in central Spain observed that land use changes determine the spatial and temporal evolution of plant cover, which directly influences trends in water resources, soil erosion and conservation. According to them, changes in land use especially in terms of vegetation composition will bring about changes in sediment production in the watershed. Also, Nadal et al. (2008) observed that a small catchment in a badland area of relatively humid environments (Mediterranean areas) shows highly active processes of physical and chemical weathering related to seasonal variations in moisture (rainfall) and temperature. Bracken and Kirkby (2005) in two semi-arid catchments of southeast Spain showed that a storm event on 20th June 2002 of 83.0mm was responsible for a maximum runoff depth of 12cm and a maximum hillslope sediment transport of 1886 cm^3/m which suggested that measured sediment transport is related to runoff. In the hilly areas of Loess Plateau, North China, Zheng et al. (2008) observed that the mean sediment concentration tends to be stable for large flood events, suggesting a strong similarity between surface flow-sediment relationships at inter and intra-event temporal scales. However, Wei et al. (2007) also in the semi-arid loess hilly area in China observed that the processes of runoff and sediment/soil loss are complicated and uncertain with the interaction of rainfall and land use which is due mainly to different stages of vegetation succession. Adriana et al. (2012) in south western France observed that grass strips along rivers and ditches prevented soil sediments from entering the surface water but did not reduce soil losses and that crop redistribution within the catchment was as efficient as planting grass strips. Other studies especially in Nigeria, have also examined the effects of both rainfall characteristics and land use/vegetation on sediment dynamics of 3rd order catchments in southwestern Nigeria. In this regard, apart from Adediji and Jeje (2004)'s work, there is little or no known studies on sediment yield dynamics from either the 1st or 2nd order basin in southwestern Nigeria in particular and in Nigeria as a whole.

Hence, this study will attempt to relate rainfall to sediment yield dynamics in a 2nd order urbanized catchment in the University of Ibadan, Ibadan, Oyo-State, Southwestern Nigeria. The 1st and 2nd order basins are ideal for the study of hydrological response pattern because they are relatively small and have homogenous physiographic and land use/vegetation attributes (Adediji and Jeje, 2004). More importantly they can respond very quickly to rainfall events in the form of storm-flow as well as to drought in sparsely vegetated areas (e.g. urbanized catchments). Hence, the main objective of this study is to relate rainfall to sediment production from a 2nd order stream draining through the built up part of University of Ibadan Estate. This study will further advance the research frontier on aspect of sediment dynamics from a small urbanized river catchment in this part of the world.

STUDY AREA

River Awba, a 2nd order river catchment within the estate of the University of Ibadan, Ibadan, Nigeria constitutes the study area. The study river basin is between Latitudes 7^{o}25'58" and 7^{o}26'42" and Longitudes 3^{o}53'21" and 3^{o}54'26" East of Greenwich Meridian (*Fig. 1*). The drainage area is 2.08 km^2, its drainage density is 1.93 km/km^2. River Awba drains through a part of the academic area of the university especially the Faculties of Science and Social Sciences as well as Departments of Petroleum and Agricultural Engineering, and emptied its water into University dam/reservoir that is very close to the Zoological Garden of the university (*Fig. 1*) The dam on R. Oba at the university has been silted up and overgrown by hydrophytes such as ferns and water weeds among which are *Pistia stratiotes*, *Scirpus cubensis* and Rhnchospora corymbosa. Other common plants around the dam include *Lemna spp; Wolffia arrhiza, Nymphaea spp;* and *Ipomoea aquatica*. Specifically, these water weeds have virtually completely colonized the surface area of the reservoir. Because, the study stream basin is mostly built-up, it usually experiences annual flooding. In fact, in August/September 2011, many properties including animals at the University Zoological Garden were destroyed by floods. Specifically, the land use map *(Fig. 2)* of the study area shows that paved/built-up area and the grass vegetation constitutes the largest proportion of the study catchment. Other land uses such as swamp and gallery vegetation along the stream channels covered the smaller portion of the study area.

The stream basin is underlain by Precambrian Basement Complex Rocks. Specifically, the area is underlain by granites, gneisses and schists (Symth and

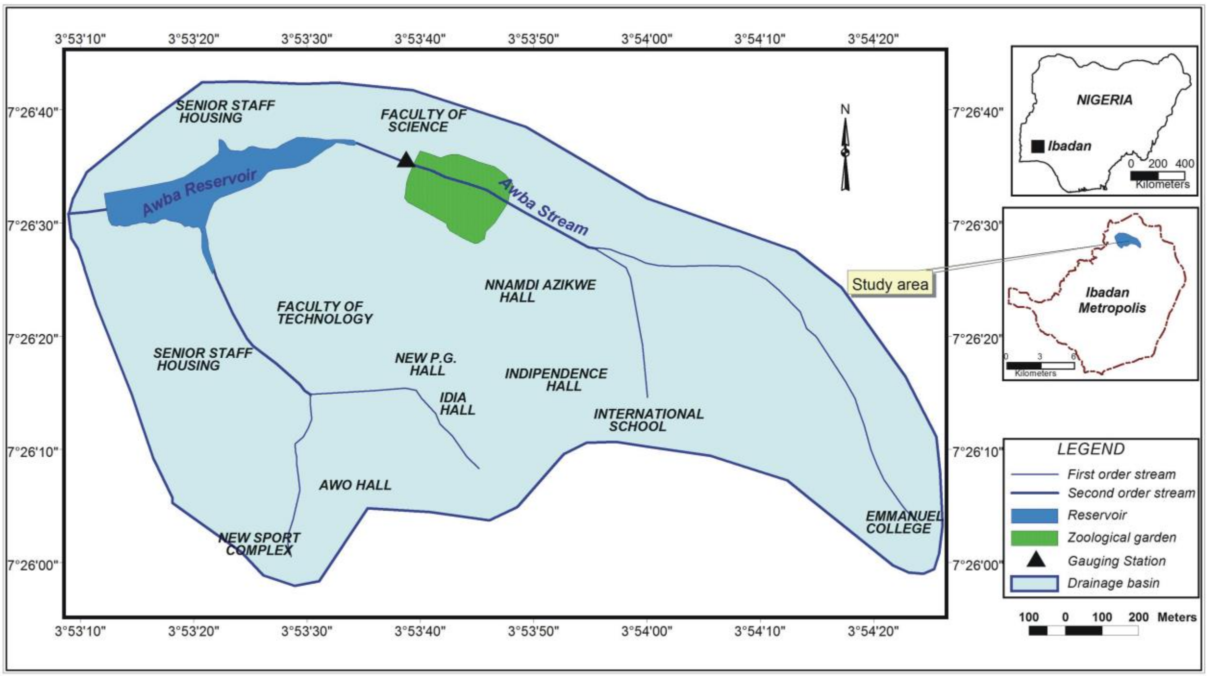

Fig. 1 Map of the Drainage Basin of River Awba within the University of Ibadan

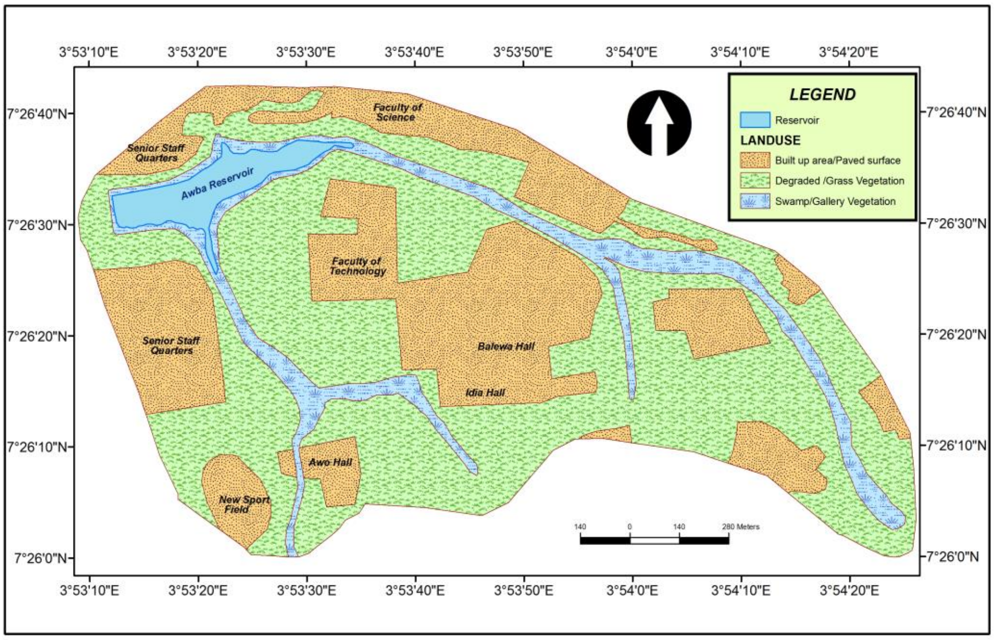

Fig. 2 Land use map of the River Awba catchment within the University of Ibadan

Montgomery, 1962). It is under Köppen's Af humid tropical rain forest climate. The mean annual rainfall is about 1400 mm and distributed between the months of March and October with peaks in July and September and a short dry spell in August although, thus varying from year to year in its occurrence (Iloeje, 1981). The rainfall effectiveness is between 6–9 months in the year. The onset and withdrawal of rains are marked by thunderstorms accompanied by high rainfall intensity. The temperature is high and almost uniform throughout the year because of the tropical climatic conditions with mean monthly value of about 27°C, while daily maximum temperatures ranges between 25°C and 30°C depending on the location and season (Iloeje, 1981).

MATERIALS AND METHODS

The study river catchment was delineated from the topographical sheet of Ibadan N.W. on a scale of 1:50,000. The topographic map used in this study was corrected following the methods suggested by Morisawa (1957) and Morgan (1971). Thereafter, the drainage network of the corrected maps was then ordered using strahler's method. Subsequently, the attributes of the study basin such as Basin area (A) and drainage density (Dd) were determined following the method by Gregory and Walling (1973).

Land use map of the study catchment was compiled from Google Image of the area at 2.5 m spatial resolution. The study stream was gauged at its exit point immediately after the Zoological Garden. The gauging station was installed at the exit point to monitor change in water level during the study period (from January 2012 until December 2012). The gauge reading was observed twice a day, in the morning (around 7.30 am) and in the evening (about 5.30 pm). The daily readings of the staff gauge were obtained for the study stream. Also, the stream flow discharge was determined using velocity-area technique. Details of the procedures involved are documented elsewhere (see Oluwatimilehin, 1991; Adediji, 2003). This was done at various stages/water levels and used to derive the discharge rating equation for the study stream. The discharge rating equation derived for the study catchment was expressed as:

$$\text{Log } Q = 0.138 + 1.062 \text{ Log } H$$

where:

Q = stream flow discharge (l/s)
H = stage/water level (cm)

The rating equation derived for the study stream was used to convert daily stage to discharge. Water sample was taken weekly during the study period. However, sampling was intensified during rainy season. In this regard, the storm runoff generated from rainfall events that occurred during the day were all sampled for the determination of suspended sediment concentration. The rating curve derived was used to obtain sediment concentration for discharges for which sampling was not done. Therefore, the rating curve technique was used to convert the stream flow discharge (m^3/s) or (l/s) to sediment concentration (mg/l) (see Miller, 1961; Walling, 1977).

Determination of suspended load involved the filtration of each 100 ml stream water sample suing Whatman Glass Fibre Circles (GFC) and a vacuum pump assembly, oven drying, cooling in a desiccator and weighing the sediment residue together with filter paper. The weight of the filter paper was subsequently subtracted to determine the weight of the residue expressed in mg/l (see Oluwatimilehin, 1991; Adediji, 2003).

Rainfall data especially daily and monthly rainfall amount between January and December 2012 was obtained from the automated weather station situated within the study catchment and under the supervision of the Department of Environmental and Agricultural Engineering, University of Ibadan, Ibadan, Nigeria.

RESULTS AND DISCUSSION

The result on the area extent covered by each of land uses identified and classified from the Google Image of the area is as shown in *Table 1*. According to *Table 1*, the largest proportion of the study basin is currently grass/degraded vegetation surface. This is distantly followed by built-up/paved surfaces and swamp/gallery vegetation at the exit point and along the stream channels, respectively. This further indicated that substantial portion of the study catchment is exposed to direct rain drop impact which might produce accelerated erosion as well as flooding. As evident from *Table 2*, the values of storm sediment concentration obtained for the study urbanized stream ranged from 636 mg/l on 21st May, 2012 to 3792 mg/l on 15th of October, 2012 with mean value yield of 2136.8 mg/l and standard deviation of 1290.9. Also, the value of storm suspended sediment yield recorded at the beginning of the rainy season especially on 4th of April 2012 (2375 mg/l) is far higher than value recorded at the middle of the wet season on the 7th of July 12 (648 mg/l) (see *Table 2*). This may not be unexpected as a lot of wastes dumped into the drainage channels and bare surfaces around the students' hostels and academic areas within the interfluvial areas of the study catchment are moved by storms runoff into the stream.

Table 1 Areal extent (m^2) of the land uses classified from the Google Earth Image of the study catchment

Land Use	Areal extent (m^2)	% of the study area
Built-up/Paved surface	702071	33.37
Swamp/Gallery Vegetation	243511	11.57
Degraded/Grass Vegetation	1158453	55.06
TOTAL	2104036	100

However, with the progressive development in the bush regrowth around the study stream channel during the peak of the rainy season coupled with high rainfall interception by bush/plant cover may possibly account for relatively low sediment concentration of 648 mg/l recorded on 7th of July, 2012. Generally, the high mean suspended sediment concentration (2136.8 mg/l) obtained for the study stream may not be unexpected because of the urbanized nature of the study catchment, where the predominant grassy vegetation is trampled as footpaths and connection route to most buildings within the studied catchment. As expected the higher mean suspended sediment concentration was high because low interception, high runoff velocity, and less time for water to infiltrate to the soil. As shown in *Table 2*, the values of storm sediment concentrations compared favourably with the results obtained from urbanized catchments in the same general area of South-western Nigeria and other parts of the humid tropical region (e.g. Oyegun, 1980 in Ibadan Northeast (Upper Ogunpa); Adediji and Jeje, 2004 in Ile-Ife; Jimoh, 2005 in Ilorin, Southwestern Nigeria; Pushparajah, 1985 in Thailand). For instance, the maximum sediment concentration (3792 mg/l) obtained for the study stream, though higher but compared favourably with the storm sediment concentration of 3475.55 mg/l recorded by Adediji and Jeje (2004) from an urbanized 2nd order stream (Odo-Ogbe) draining Oja-Titun area in Ile-Ife, Southwestern Nigeria. Also, the highest value obtained in this study is in accordance with the maximum value of 4780 mg/l obtained by Pushparajah (1985) from Huay Ma Feang Stream (urbanized streams) in Thailand during the rainy season of 1983. Also, the values of maximum suspended sediment concentration obtained is in accordance with the findings by Fang et al. (2011) in a small agricultural watershed of the three Gorges in China where maximum storm flow sediment concentration varied from 183 to 62, 138g/m^3 with a mean suspended sediment concentration of 7962g/m^3.

Further, as shown in *Table 2*, the monthly total sediment yield for the study streams ranged from 10.85kg in January to 288.40 kg in October with the mean soil loss of 123.59 kg and standard deviation of 89.50. The monthly suspended sediment yield rose from 10.85 kg in April to 262.70 kg in July and declined to 107.58 kg in September and rose again to 288.40 kg in October and eventually decreased to 18.75 kg in December, 2012 more or less synchronously with the monthly rainfall. It is quite evident from *Fig. 3* and *Table 2* that the monthly total sediment yield closely follows the pattern of monthly rainfall. This was also in accordance with the observation made by Adediji et al (1995) in Ile-Ife area of South-western Nigeria.

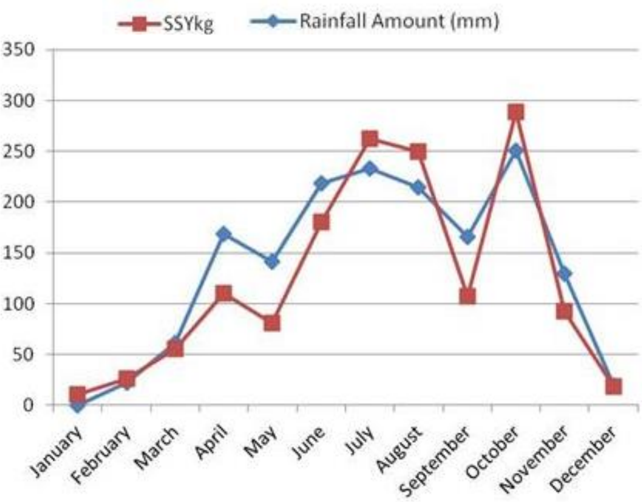

Fig. 3 Relationship between amount of rainfall (mm) and the sediment yield of the River Awba catchment

Table 2 Rainfall events and suspended sediment concentrations of the study catchment

Rainfall Event	Date	Rainfall Amount (mm)	Suspended Sediment Load (mg/l)[a,b]
1	4-4-12	32.7	2375
2	20-4-12	32.7	1053
3	4-5-12	32.7	643
4	21-5-12	32.7	636
5	7-7-12	44.1	648
6	15-7-12	111.9	3308
7	16-7-12	122.7	3721
8	20-8-12	30.3	979
9	20-9-12	78.3	3641
10	25-9-12	70.9	2350
11	10-10-12	101.7	2494
12	15-10-12	154	3792

[a]Mean suspended sediment load = 2136.8mg/l
[b]Standard deviation of suspended sediment load = 1290.9

As evident from Fig. 4, the storm suspended sediment concentrations significantly related to storm flow discharge of the study stream ($r = 0.71$ at $p = 0.01$). This further confirmed the findings by Nadal et al. (2008) in a small catchment with badlands in Spain where significant relationships was obtained among rainfall, runoff and suspended sediment. The result obtained in this study is also in accordance with the findings of Zokaib and Naser (2012) in Hilkot watershed of Pakistan where a good relation was observed between rainfall, runoff and soil/sediment loss under different land uses.

CONCLUSIONS

The dynamics of suspended sediment yield in a 2nd order River Awba stream within the estate of University of Ibadan, Ibadan, Nigeria was undertaken in this study. The study was carried out between January and December, 2012. The results showed that storm suspended sediment concentration was relatively higher at the beginning of the rainy season (2375 mg/l (4th of April, 2012) than at the middle of wet season (648 mg/l) (7th of July, 2012). However, the monthly suspended increases with the increase in monthly rainfall amount. For instance, the monthly sediment yield increase by 240.25 kg between January and October. The total sediment load discharged from the study river catchment was estimated at 1.48 tonnes/year which compared favourably with Adediji and Jeje's (2004) findings from the same general study area (i.e. Southwestern Nigeria).

In the light of the above, in order to minimize the rate of storm runoff and load/yield generated from rain storm events in the study area, the paved surfaces in the interfluvial areas of the study river catchment should be grassed and avoid trampling .The inclusion of rapid growing tree species will enhance the reduction in the sediment loss into the study stream. This will subsequently minimize the rate of flood generation around the Zoological garden of the university in the study catchment. This is in accordance with the observation made by Adriana et al. (2012) in southwestern France that in order to preserve the quality of surface water as well as reduction of sediment concentration, the farmers should keep a minimum acreage of grass land especially in areas bordering the river channel as also required by official French regulation.

Acknowledgements

We appreciate the sincere efforts of Mr. Adigun of the Remote sensing laboratory of the Department of Geography, University of Ibadan for his prompt and adequate

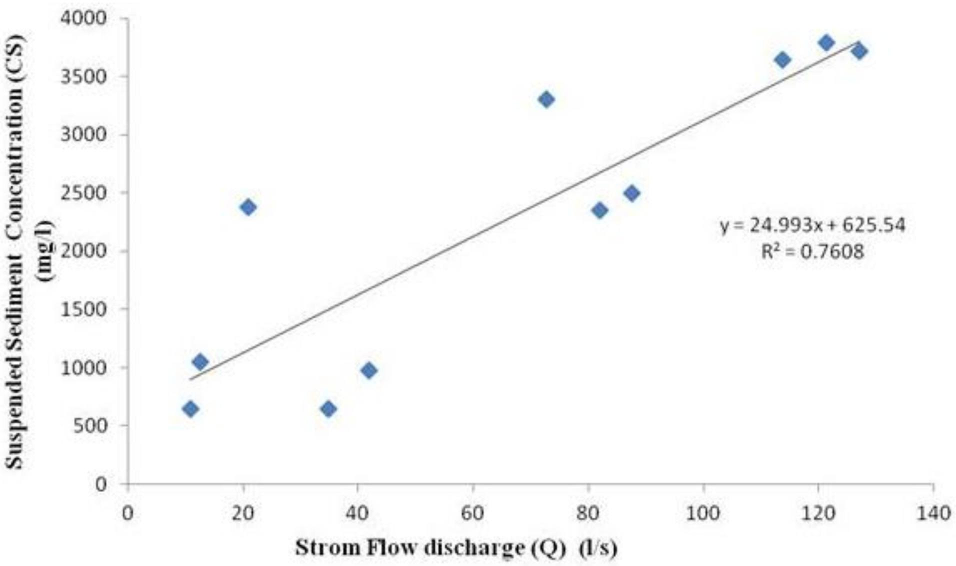

Fig. 4 Relationship between suspended sediment (CS) and stream flow discharge (I/s) (Q)

measurements of the stage during the study period and also Mr. Olofinyo of the Physical Geography Laboratory for analysing the sediments.

References

Adediji, A. 2003. Sediment delivery ratios in Opa Reservoir Catchment, Southwestern Nigeria, Unpublished Ph.D Thesis, Department of Geography, Obafemi Awolowo University, Ile-Ife, Nigeria.

Adediji, A., Jeje, L.K. 2004. Channel erosion in the Opa Basin, Southwestern Nigeria. *Journal of Environmental Hydrology* 12, 1–11.

Adediji, A, Jeje, L.K., Ibitoye, M.O. 2013 Urban development and informal drainage patterns: Gully dynamics in Southwestern Nigeria, *Applied Geography* 40, 90–102. DOI: 10.1016/j.apgeog.2013.01.012

Adriana, F, Jean-Christophe, P, Jean-Claude, M., Yveg, L.B. 2012. Designing management options to reduce surface runoff and sediment yield with Farmers: An experiment in south-western France. *Journal of Environmental Management* 96, 74–85. DOI: 10.1016/j.jenvman.2011.11.001

Bracken, L.J., Kirkby, M.J. (2005): Differences in hillslope runoff and sediment transport rates within two semi-arid catchments in Southeast Spain, *Geomorphology*, 68, 183–200. DOI: 10.1016/j.geomorph.2004.11.013

Fang, F.N., Zhi, H.S., Lu, L., Zhang, G., Gian-Jin, L., Lei, A. 2012. The effects of rainfall regimes and land use changes on run off and soil loss in a small mountainous watershed, *Catena* 99, 1–8. DOI: 10.1016/j.catena.2012.07.004

Fang, N., Shi, Z., Li, L., Jang, C. 2011. Rainfall, runoff, and suspended sediment delivery relationships in a small agricultural watershed of the Three Gorges area, China. *Geomorphology* 135, 558–568. DOI: 10.1016/j.geomorph.2011.08.013

Gao, P. 2008 Understanding watershed suspended sediment transport. *Progress in Physical Geography* 32, 243–248. DOI: doi: 10.1177/0309133308094849

Gregory, K.J. and Walling, D.E. (1973) Drainage Basin, Form and Process. A Geomorphological Approach. Edward Arnold, London, 58 pp.

Jeje, L.K, Ogunkoya, O.O., Oluwatimilehin, J.M. 1999. Solute load concentrations in some streams in the Upper Osun and Owena drainage basin, central western Nigeria. *Journal of African Earth Science* 29 (4), 799–808. DOI: 10.1016/s0899-5362(99)00130-x

Jimoh, H.I. (2005) Tropical rainfall events on erosion rates in a rapidly developing urban area in Nigeria. *Singapore Journal of Tropical Geography* 26, 74–81. DOI: 10.1111/j.0129-7619.2005.00205.x

Iloeje, N.P. 1981. A new geography of Nigeria, new revised edition. Longman Publishers, London.

Mwamba, M.J., Torres, R. 2009. Rainfall effects on marsh sediment redistribution, North Inlet, South Carolina, USA. *Marine Geology* 89, 267–287. DOI: 10.1016/s0025-3227(02)00482-6

Miller, J.P. 1961. Solutes in small streams draining single roc k types San gre de Cristo Range, New Mexico, U.S. Geol. Survey Water Supply Paper 1535F.

Morgan, J.P. 1971. Soil Erosion. Longman, London and New York.

Morisawa, M.E. 1957. Accuracy of determination of stream length from Topographic Maps Trans. Amer. *Geophysical Union* 38, 86–88. DOI: 10.1029/tr038i001p00086

Nadal-Romero, E., Regues, D. 2008. Sediment yield from badland areas in Mediterranean environments. *Progress in Physical Geography* 29, 182–190.

Nadal-Romero, E., Regnes, D., Latron, J. 2008. Relationships among rainfall, runoff and suspended sediment in a small catchment with badlands. *Catena*, 74, 127–136. DOI: 10.1016/j.catena.2008.03.014

Slattery, M.C., Phillips, J.D. 2010. Controls on sediment delivery in coastal plain rivers. *Journal of Environmental Management* 92, 284–289. DOI: 10.1016/j.jenvman.2009.08.022

Nu-Fang, F., zhi-Hua, S., Lu, L., Zhong, G., Lei, A. 2012. The influence of rainfall regime and land use on runoff and soil loss in a small mountainous watershed. *Catena* 99, 1–8. DOI: 10.1016/j.catena.2012.07.004

Pushparajah, E. 1985 Development of induced soil erosion and flash floods in Malaysia. *Ecologist* 15 (1-2), 19–25.

Smets, T., Poesen, J., Bochet, E. 2008. Impact of plot length on the effectiveness of different soil surface covers in reducing runoff and soil loss by water. *Progress in Physical Geography*, 32, 654–661. DOI: 10.1177/0309133308101473

Taguas, E.V., Ayuso, J.L., Perez, R., Giraldez, J.V., Gomez, J.A. 2013. Intra and inter-annual variability of runoff and sediment yield of an Ohre micro catchment with soil protection by natural ground cover in southern Spain. *Geoderma*, 206, 49–62. DOI: 10.1016/j.geoderma.2013.04.011

Thair, O., Anas, D. 2010. River-basin planning and management. *Geoforum* 40, 484–494.

Walling, D.E. 1977. Assessing the accuracy of suspended sediment rating curve for a small basin. *Water Resources Research* 13(3), 531–538. DOI: 10.1029/wr013i003p00531

Wei, W., Liding, C., Bojie, F., Zhilin, H., Dongping, W. 2007. The effect of land uses and rainfall regimes on runoff and soil erosion in the semi-arid loess hilly area, China. *Journal of Hydrology* 335, 247–258. DOI: 10.1016/j.jhydrol.2006.11.016

Zheng, M., Qiangguo, C., Qinjua, C. 2008. Modelling the runoff-sediment yield relationship using a proportional function in hilly areas of Loess Plateau, North China. *Geomorphology* 93, 288–301. DOI: 10.1016/j.geomorph.2007.03.001

Zokaib, S., Naser, G. 2012. A study on rainfall, runoff and soil loss relations at different land uses. A case in Hilkot watershed in Pakistan. *International Journal of Sediment Research* 27, 388–393. DOI: 10.1016/S1001-6279(12)60043-2

JOURNAL OF ENVIRONMENTAL GEOGRAPHY
Journal of Environmental Geography 7 (1–2), 29–38.
DOI: 10.2478/jengeo-2014-0004
ISSN: 2060-467X

AUTOCHTHONOUS VERSUS ALLOCHTHONOUS ORGANIC MATTER IN RECENT SOIL C ACCUMULATION ALONG A FLOODPLAIN BIOGEOMORPHIC GRADIENT: AN EXPLORATORY STUDY

Eduardo González [1, 2*], Álvaro Cabezas[3], Dov Corenblit[4, 5], Johannes Steiger[4, 5]

[1] UPS, INP, EcoLab, Université de Toulouse, 118 Route de Narbonne Bâtiment 4R1, F-31062 Toulouse Cedex 9, France
[2] CNRS, EcoLab, F-31062 Toulouse Cedex 9, France
[3] Chemical Analytics and Biogeochemistry, Leibniz Institute of Freshwater Ecosystems and Inland Fisheries, Müggelseedamm 301, 12587, Berlin, Germany
[4] Clermont Université, UBP, MSH, 4 rue Ledru, 63057 Clermont-Ferrand Cedex 1, France
[5] CNRS, UMR GEOLAB – Laboratoire de géographie physique et environnementale, 63057 Clermont-Ferrand, France
*Corresponding author, e-mail: edusargas@hotmail.com

Research article, received 25 February 2014, accepted 30 March 2014

Abstract

The mechanisms controlling soil succession in floodplains remain much less studied than in uplands due to the complexity that flood-driven erosion and sedimentation bring into soil development processes. The amount of organic matter and C generally grows with soil ageing and is controlled by multiple and interacting allogenic and autogenic factors, but to what extent the production of organic matter by *in situ* vegetation contributes to soil formation in floodplains remains unknown. The objective of this work was to explore the importance of autochthonous organic matter versus allochthonous organic matter in organic C accumulation of floodplain forest soils along a vegetation succession and hydrogeomorphic connectivity gradient. Physicochemical analyses of sediment collected after one single flood event in a large Mediterranean floodplain (Middle Ebro, a 9^{th} order regulated river reach in NE Spain) were used to estimate the proportion of organic C found in the topsoil (first 10 cm) samples of young (<25 yr), mature (25-50 yr) and old (>50 yr) floodplain forests that had an allochthonous (i.e., % of organic C deposited by floods) or autochthonous (i.e., % of organic C produced *in situ* by vegetation) source. Results of this exploratory study showed that the accumulation of autochthonous organic C in the floodplain topsoil only occurred in floodplain forests older than 50 year-old, but even then, it was more than six-fold less abundant than that with an allochthonous origin. Moreover, a linear mixed effect model showed that, although autochthonous organic C accumulation was mainly explained by the forest structure, a small proportion of it was also controlled by an allogenic factor, the groundwater table depth. Then, groundwater table depth variations could be partly controlling autochthonous organic matter production and decomposition in this Mediterranean floodplain. Although flow regulation and embankment has dramatically limited the hydrogeomorphic dynamics of the river, allogenic overbank sedimentation during flood events still controls floodplain soil succession and organic C accumulation in the floodplain.

Keywords: floodplain soils, sediment texture, soil organic carbon, overbank sedimentation

INTRODUCTION

Organic matter accumulation within floodplains is an ecosystem function that provides important benefits such as water quality enhancement or mitigation of greenhouse effect (Johnston, 1991; Day et al., 2004; Verhoeven et al., 2006; IPCC, 2007). Soil organic matter is especially relevant since it forms a larger and more lasting pool than living biomass. However, soil successional patterns have been much less studied in floodplains than in upland forests (Wigginton et al., 2000). The main reason is that successional trajectories in floodplains are recurrently brought to a starting point by destructive floods (Hughes et al., 1997; Geerling et al., 2006; Cabezas et al., 2009a), thus making the forest floor development much more unpredictable than in the more stable uplands.

The role of floodplain soils as sink or source of organic C varies along the longitudinal river axis, with downstream river sections importing the suspended sediment and a part of organic matter of the upstream sections of the catchment (Steiger and Gurnell, 2003; Noe and Hupp, 2005). Along the transversal river axis, from the water channel to the floodplain, flood events determine the role of a given ecotope as sink or source of organic matter as well as its source (allochtonous *vs.* autochthonous) for a given study period. Organic matter produced by biotic assemblages is incorporated into surface soils during the intervals between overbank flooding. During floods, low-organic fluvial sediments may be deposited on floodplain landforms, and the organic matter that had been stored in different compartments such as the soil or standing vegetation is either exported or buried (Cabezas and Comín, 2010), maintaining the soil in a continuous immature state (Drouin et al., 2011). As vegetation develops in the absence of destructive floods, the floodplain surface roughness increases and enhances sediment deposition, which in turn causes its elevation in relation to the river channel to rise (Hupp, 2000; Bendix and Hupp, 2000; Steiger et al., 2001). Sandy sediment trapping by young and dense thickets of vegetation predominates in the most exposed zones, while organic matter produced *in situ* may be more easily retained in less exposed zones such as the hydrologically disconnected floodplain.

Overbank sedimentation of fine sediment especially favors the accumulation of organic C in the soil as organic matter forms easily complexes with fine sediment (silt and clay <63 µm) (Asselman and Middelkoop, 1995; Steiger and Gurnell, 2003), silt and clay fraction and organic carbon being positively correlated (Pinay et al., 1992; Bechtold and Naiman, 2006). However, accretion is also negatively feedbacked by hydrological disconnection (Bendix and Hupp, 2000). Both increasing distance to and elevation above the main channel generally contribute to the decrease in duration and frequency of flooding (Corenblit et al., 2007), eventually leading to a decrease in the input of sediment and organic matter bounded to sediment particles (i.e., allochthonous organic matter) (Asselman and Middlekoop, 1995; Walling and He, 1997; Piégay et al., 2008; Cabezas et al., 2010). In other words, the input of organic C in a given forest patch via sedimentation is limited to areas that are hydrogeomorphologically connected to the main channel. On the other hand, as floodplain forest succession and hydrogeomorphic disconnection develop, litter (i.e., authochthonous organic matter) accumulates in the topsoil because it is less likely to be intensively and frequently removed by floods (Friedman et al., 1996). It follows that similar to floodplain vegetation dynamics (Corenblit et al., 2007), the soil development and organic C accumulation is driven mainly by abiotic factors at the early stages of soil succession, implicating that during the allogenic vegetation succession stage allochthonous organic matter dominates. But if these hydrogeomorphic drivers are gradually replaced by biotic factors in later stages of succession, i.e. late-seral stages, it is expected that during autogenic succession stages autochthonous organic matter dominates. This theoretical framework of soil succession would fit with classic succession theory (Clements, 1916). However, studies devoted to quantify these processes within floodplain zones have not been frequent, and the relative roles of abiotic (hydrology, topography and sediment dynamics) and biotic (forest structure) factors as drivers of organic C dynamics in floodplain soil development remain largely unknown (but see Cierjacks et al., 2011). More particularly, although it is widely recognized that floodplains are highly productive ecosystems (Naiman et al., 2005), little is known about how much autochthonous sources contribute to organic C accumulation locally. In fact, the efforts done to understand patterns of organic C accumulation in floodplain soils have rarely considered autochthonous and allochthonous sources separately (but see Daniels, 2003; Cabezas and Comín, 2010).

The objective of this study was to evaluate the relative importance of autochthonous *vs*. allochthonous organic C (respectively AUTOC and ALLOC) along a vegetation succession and hydrogeomorphic connectivity gradient (a biogeomorphic gradient *sensu* Corenblit et al. 2007) in the soil of a large forested floodplain. ALLOC was expected to be more important at the more connected and younger sites, with abiotic factors driving succession (Corenblit et al., 2007) and AUTOC higher at disconnected and older forest patches, with biotic factors driving succession (Corenblit et al., 2007).

STUDY AREA

The study was carried out in an 8 km river segment in the Middle Ebro River. The segment was located 12 km downstream from the city of Zaragoza (41º36' N, 0º46' W, 175-185 m a.s.l., NE Spain). The Ebro River is one of the largest Mediterranean rivers in terms of length (~930 km), annual discharge (~12 000 $hm^3\ yr^{-1}$) and drainage area (~86 000 km^2). The hydrological regime is pluvio-nival and the river is highly regulated by dams built for irrigation and flood control during the 1950's and 1960's. At the river gauge station of Zaragoza, the annual mean of the daily flow discharge between 1981 and 2003 is 230 $m^3\ s^{-1}$. As a consequence of flow regulation, a decrease in bankfull discharge and in the duration and frequency of overbank flooding events throughout the second half of the 20th century was observed (Cabezas et al., 2009a). Both the selected river segment and the middle reach of the Ebro have a very limited channel migration rate since the 1970's and 1980's, due to two decades of intense bank protection works (Ollero, 2007; Magdaleno and Fernández, 2011).

However, the middle reach of the Ebro still maintains a sinuous shape (sinuosity = 1.39, mean longitudinal channel slope = 0.05 %, Ollero, 1995) as a relict of the former dynamic geomorphic regime before river training works. The average width of the submersible floodplain is 5 km but agriculture and human settlements have reduced the surface occupied by natural floodplain forest to ~4.5% of the floodplain surface (Ollero, 2007), usually occupying the inside of meander bends, which are not protected by bank protection works. Despite protection by levees, floodplain forests on both the convex and concave river banks are inundated by floods with recurrence intervals of <10 years (1981-2003). Tree communities are dominated by phreatophytic species (*Populus alba* L., *Populus nigra* L., *Salix alba* L., *Tamarix gallica* L., *T. africana* Poir. and *Tamarix canariensis* Willd.). Late-seral hardwood species (*Ulmus minor* auct. non P. Mill. and *Fraxinus angustifolia* Vahl) are commonly found, but usually with size stems smaller than 7.5 cm d.b.h. (González et al., 2010a).

METHODS

Plot selection along a vegetation succession and hydrogeomorphic gradient

Using a series of aerial photographs (1957, 1981 and 2003), 39 forest patches of different age: young (<25 years, 13 patches), mature (25-50 years, 13 patches) and old (>50 years, 13 patches), all subjected to annual flooding along a hydrogeomorphic gradient, were selected for this study (*Fig. 1*).

One rectangular study plot was randomly placed within each forest patch to carry out a tree survey in 2006 and 2007, including the diameter and species record of every woody stem ≥ 0.3 m height (total = 6891 stems) (González et al., 2010a). Plot dimensions were scaled according to the dominant tree height and ranged from 25 to 1000 m^2. The collected information was used to calculate total and species-specific basal areas (BA, m^2 ha^{-1}) and stem densities (SD, stems ha^{-1}) following González et al. (2010a). Species-specific importance values (IV) were then calculated at each plot using a modification of the Gergel et al. (2002) formula, ranging from 0 (absence) to 1 (complete dominance): IV_i = [(SD of species i / total SD) + (BA of species i / total BA)] / 2.

To describe the hydrogeomorphic regime at each plot, a network of 17 piezometers was installed in the floodplain (details in González et al., 2012). Water levels in the piezometers were continuously monitored from October 2008 to September 2010 (highest flood peak recorded at the gauging station of Zaragoza during the monitoring period: 1506 m^3 s^{-1}, February 2009, recurrence interval for the 1927-2003 period: 0.87 yrs) using integrated absolute pressure sensors and data loggers (van Essen DI502 TD Diver®), which permit the determination of groundwater table depths as well as water stages above piezometers during inundation and thus above the floodplain surface. Hereafter the synonym ‘water table’ was used for groundwater table and ‘floodwater stage’ for water stage above piezometers during inundation. The piezometer measurements were used to interpolate the water table levels and floodwater stages and, following González et al. (2012), to determine the deepest water table level recorded (WT, m), flood duration (FD, % of the time) and flood frequency (FF, number of flood submersions) that represented the local hydrologic regime at each plot during the study period. The piezometer and plot elevations above sea level (ELEV, m) and the UTM coordinates necessary for all calculations had previously been recorded using a differential global positioning system (DGPS) Topcon® with 3 cm vertical accuracy. The shortest distance from the centroid of each plot to the main river channel at the summer level (DIST, m) was calculated using ArcGis 9.2.

Topsoil collection and processing

Soil samples were taken at the end of the vegetation growing season in September and October 2006, coinciding with low water levels prior to autumn floods and before the main peak of litterfall that typically occurs in November in the study area (González, 2012a). This sampling period was chosen to minimize the influence of the litter fall over the soil organic matter analyses, thus focusing on the consolidated organic matter soil pool. At each plot, three topsoil (0–10 cm) samples were collected using an undisturbed soil sampler (5 cm diameter steel tube, P.1.31 Eijkelkamp®) after carefully removing the living vegetation and litter layers. At the larger plots (>500 m^2, 6 plots), more samples were collected to obtain a composite of three samples that better represented the heterogeneity of the soil in the plot. Samples were air dried before being passed through a 2 mm sieve to remove rock and larger organic matter fragments. Total organic C (TOC, %) was measured with a LECO SC 144 DR® elemental analyzer. The fine fraction (i.e., FINE silt + clay, % of particles <63 μm) was calculated gravimetrically by sieving sub-samples

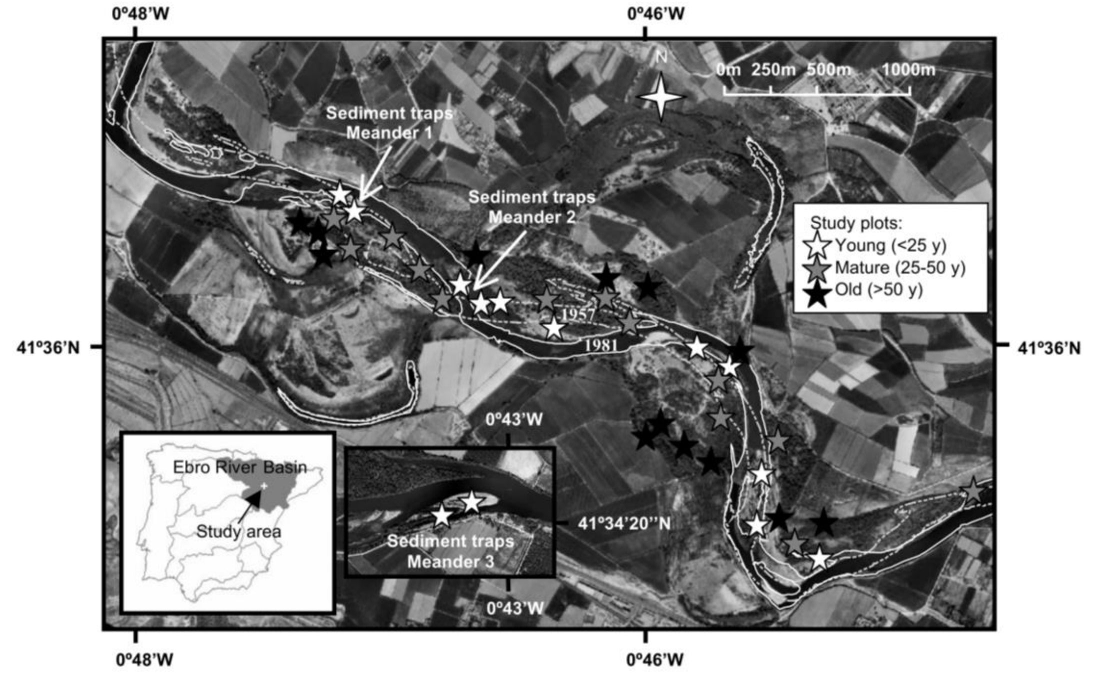

Fig. 1 Location of the study sites in a 8 km segment of the Ebro River. The aerial picture was taken in 2007. Lines correspond to the main channel tracks in 1957 and 1981

of the < 2 mm material into a 63 μm screen previously treated with a 10% hydrogen peroxide solution to eliminate the organic matter and with a polyphosphate solution to facilitate particle dispersion. In young plots, a framework of gravels, cobbles and/or boulders was usually present in the surface layer and the soil sampler could not be used. Then, the soil analyses were performed using the finer sediments filling the pore spaces between the framework grains.

Determination of ALLOC and AUTOC

The particle-size composition of sediments partially determines the amount of organic C that is deposited locally during floods. The sediment deposited during one flood event was collected using 92 sediment traps made of artificial grass mats (25 x 25 cm) set up in 3 meanders of the study area (*Fig. 1*). Details on the experimental setup can be found in Cabezas et al. (2010). Artificial grass mats have been commonly used to study contemporary sedimentation rates in river floodplains, especially when it was necessary to recover the sediment deposited during one individual flood event as in the present study (see Steiger et al. 2003 for a review). The flood occurred in January 2006 and lasted for 8 days. It had a flood peak at the gauging station of Zaragoza of 754 m^3 s^{-1} and a recurrence interval for the 1927-2003 period of 0.23 yrs. A few days after the flood, mats re-emerged with only 7% of them having been flushed away by the river. Mats were collected, brought to the laboratory and oven-dried at 60 ºC. The deposited sediment was then carefully removed by hand using a metallic brush and their proportion of fine particles and TOC determined as described above for the topsoil samples. A polynomial function was adjusted to model the relationship between the fine fraction and TOC measured in the sediment gathered from the mats. The function served to calculate the theoretical proportion of TOC that was deposited during floods at each topsoil sample in the 39 study plots (i.e., ALLOC), using the fine fraction as independent variable. In other words, it was considered that the ALLOC within a given topsoil sample could be predicted as a function of its grain size (Pinay et al., 1992; Bechtold and Naiman, 2006; Cabezas and Comín, 2010). It follows that the difference between observed and predicted TOC values (TOC residuals) would be an indicator of the organic C produced and accumulated *in situ* at each topsoil sample (i.e., AUTOC). Our objective was not to calculate *absolute* organic C accumulation rates, which would have also required to measure local sedimentation rates, which was not done here. Instead, the aim was to obtain a quantitative estimate of the accumulated organic C that had been produced *in situ* (TOC residual = AUTOC), compared to the accumulated organic C deposited by floods (predicted TOC = ALLOC). Negative values in the TOC residuals of some plots were an unavoidable artifact of our experimental approach and were explained by the fact that the floods that were actually responsible for the ALLOC deposition in those plots contained a lower organic content than the flood used to build the predictive function. However, negative TOC residuals were interpreted as indicators of a lower probability of having AUTOC accumulation (or as a higher certitude in the absence of AUTOC accumulation) and were included in the analyses explained below.

Evaluating the relative importance of AUTOC vs. ALLOC along the biogeomorphic gradient

In a first step, predicted and observed TOC values were plotted together grouped by plot age and, within each age category, their medians compared by means of nonparametric Wilcoxon matched-pairs signed-ranks tests. Therefore, it was examined whether erosion-sedimentation processes controlled only the TOC topsoil levels (i.e., that if predicted and observed TOC values were not significantly different, no AUTOC was accumulated) or whether an accumulation of AUTOC effectively occurred (i.e., that observed TOC values were significantly higher than the predicted TOC values) along the floodplain forest chronosequence. The significance levels were set up at 0.05.

Statistical modeling of AUTOC accumulation

Once evaluated the relative importance of AUTOC *vs*. ALLOC along the biogeomorphic gradient, the influence of abiotic and biotic factors on the variability of the TOC residuals was assessed using linear mixed effect models (LME, Pinheiro and Bates, 2000) and hierarchical partitioning. LME are a type of multiple regression that do not assume that all observations are independent from each other, and so can be used to analyze data from clustered experimental designs where observed subjects are nested within larger units. Different LME models were run using all the variables representing the vegetation succession and hydrological connectivity (*Table 1*) as fixed factors, plot as random effect, and TOC residuals as the dependent variable, with n = 117 (39 forest plots x 3 topsoil replicates). A backward selection of predictors was performed to retain only the most significant ($P < 0.05$) explaining the response of TOC residual, testing each step with a likelihood ratio-test (Crawley, 2002). Pearson's product–moment correlation between the observed and predicted values and the Akaike Information Criterion were used to assess the goodness-of-fit of the final models. Once identified the main predictors, the sum of independent and shared variance in the TOC residuals explained by each significant predictor (i.e., the relative contribution) in each model was assessed through hierarchical partitioning (Chevan and Sutherland, 1991). The LME models and hierarchical partitioning were run using functions available in the package "nlme" (Pinheiro et al., 2007) and "hier.part" (Walsh and McNally, 2007) of the R 2.14.0 software (R Development Core Team 2011). Different transformations were applied to the explanatory variables to correct for the effect of asymmetry, given that most of them were not normally distributed (Kolmogorov-Smirnov test). The exploratory analyses showed that TOC residuals had a strong positive and heteroscedastic relationship with "observed TOC" (i.e., higher TOC residuals and higher variance were observed at higher TOC values). To remove that confounding effect, "observed TOC" was included in the models as covariable. All variables were standardized before running the models.

RESULTS

Forest succession and hydrogeomorphic gradient

The classification of plots into three age-groups along a forest succession and hydrological connectivity gradient fitted with their forest structure and with their local hydrological connectivity descriptors, with significant differences between young, mature and old plots in all variables reported (*Table 1*). Young plots were dominated by pioneer tree species *Populus nigra*, *Tamarix* spp. and *Salix alba* (sum of their IV = 0.99) and were in a very active recruitment phase, as shown by their significantly higher stem density and lower basal area compared to mature and old plots. This reflected the prevalence of small stems in the forest structure. Compared to the older plots, young plots also displayed the highest hydrological connectivity, with a significantly higher water table, flood duration and frequency as direct consequences of their lower topographical position and shorter distance to the main channel. Mature plots represented an intermediate successional stage along the chronosequence. Pioneer species were still dominating (sum of their IV = 0.81) but formed less dense patches than in young plots (stem density decreased eight-fold while basal area slightly increased). In addition, late-seral stage species, namely *Ulmus minor* and *Fraxinus angustifolia*, were now more frequent. The higher elevation plots in the floodplain were located at a greater distance from the main channel and were three-fold less flooded (in terms of duration) than young plots. Within the mature plots, the lowest water tables recorded were 1 m deeper on average than in the young plots, but flood events were still frequent. In the old plots pioneer species did not dominate (sum of their IV = 0.44), although not only as a result of their replacement by late-seral species (IV = 0.25) but also caused by the appearance of a new successional pathway dominated by *Populus alba* (present in 7 out of the 13 old plots surveyed). Old plots were the hydrologically most disconnected sites along the chronosequence, being rarely flooded and with lowest water tables recorded reaching a mean depth of ~3.5 m below the floodplain surface.

Topsoil

Overall, the soil descriptors also followed the biogeomorphic gradient (*Table 1*). Sandy soils predominated in young sites. However, mature and old topsoil forests

Table 1 Summary of the forest structure and hydrological connectivity descriptors for 39 forest patches divided into three age categories: young, mature and old. Values are means ± 1 SE. Letters indicate homogeneous groups after independent t tests implemented in SPSS v 13.0 (P < 0.05)

	Young	Mature	Old
N	13	13	13
Age (years)	<25	25-50	>50
Plot size (m^2)	$130^b \pm 33$	$386^a \pm 21$	$545^a \pm 86$
Forest structure			
Basal area (BA, $m^2\ ha^{-1}$)	$34^b \pm 8$	$48^{ab} \pm 7$	$67^a \pm 11$
Stem density (SD, stems ha^{-1})	$26395^a \pm 8365$	$3270^b \pm 749$	$5771^b \pm 1178$
IV (unitless)			
Populus nigra (IVPn)	$0.48^a \pm 0.10$	$0.35^{ab} \pm 0.07$	$0.18^b \pm 0.07$
Populus alba (IVPa)	$0.00^b \pm 0.00$	$0.00^b \pm 0.00$	$0.30^a \pm 0.10$
Tamarix spp. (IVTx)	$0.42^a \pm 0.10$	$0.29^a \pm 0.08$	$0.24^a \pm 0.08$
Salix alba (IVSa)	$0.09^a \pm 0.03$	$0.17^a \pm 0.08$	$0.02^a \pm 0.01$
Fraxinus angustifolia (IVFx)	$0.01^b \pm 0.00$	$0.11^a \pm 0.03$	$0.03^b \pm 0.01$
Ulmus minor (IVUm)	$0.00^c \pm 0.00$	$0.07^b \pm 0.02$	$0.21^a \pm 0.05$
Other species (IVOt)	$0.00^b \pm 0.00$	$0.01^{ab} \pm 0.00$	$0.01^a \pm 0.00$
Hydrological connectivity			
Maximum depth to water table (WT, m)	$1.28^c \pm 0.15$	$2.25^b \pm 0.19$	$3.49^a \pm 0.21$
Flood duration (FD, % of time)	$24^a \pm 4$	$9^b \pm 3$	$2^c \pm 1$
Flood frequency (FF, events y^{-1})	$5^a \pm 1$	$4^b \pm 0$	$2^c \pm 0$
Topography			
Elevation a.s.l. (ELEV, m)	$178^b \pm 1$	$180^a \pm 0$	$181^a \pm 0$
Distance to river channel (DIST, m)	$42^c \pm 10$	$117^b \pm 22$	$221^a \pm 37$
Topsoil (first 10 cm)			
Total organic carbon (TOC, %)	$1.22^c \pm 0.16$	$2.28^b \pm 0.11$	$2.73^a \pm 0.18$
Fine fraction (FINE, %)	$47^b \pm 7$	$88^a \pm 3$	$86^a \pm 5$

were basically composed of fine sediment. TOC progressively increased along the chronosequence starting at young sites near the river channel.

Determination of ALLOC and AUTOC

The fine-texture of the sediment deposited on the mats by the studied flood event was positively correlated with the associated TOC deposited following a quadratic relationship (*Fig. 2*). The polynomial function predicted well the proportion of organic C contained in the topsoil samples of the young and mature forests (no significant differences were found between predicted and observed TOC median values after Wilcoxon tests, *Fig. 3*).

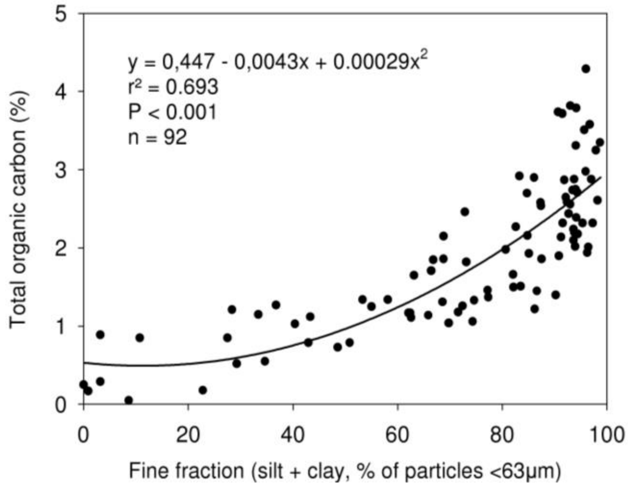

Fig. 2 Scatter plot of sediment texture (silt plus clay, %) and Total Organic Carbon (TOC, %) in the sediments deposited during a flood on 1 January 2006 in the study area

That means that all the organic C had an allochthonous origin and no AUTOC was being accumulated in the plots < 50 year-old. On the contrary, the function under-estimated the proportion of organic C that was found in the topsoil of the old age-forest category (> 50 year-old). That is, the observed TOC median was significantly higher than the predicted TOC median (*Fig. 3*). This fact was interpreted as a proof of a significant accumulation of AUTOC. However, even in this situation, the contribution of the autochthonous fraction to the organic C stock was more than six-fold lower than that of the allochthonous origin (median TOC residual in old plots, AUTOC = 0.41%; median predicted TOC in old plots, ALLOC = 2.64%).

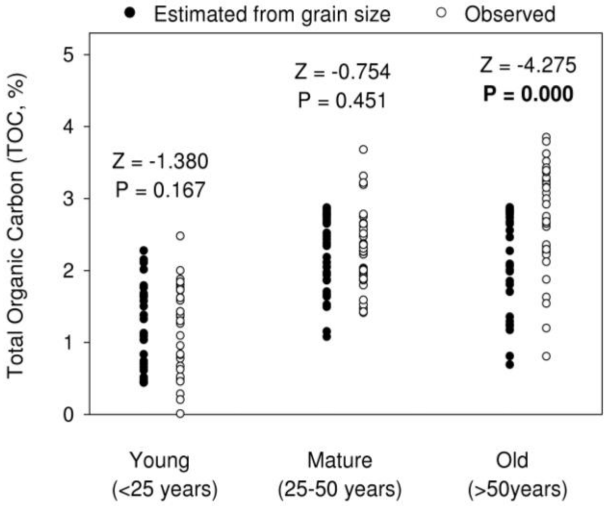

Fig 3 Observed and predicted TOC values from Equation in *Fig. 2* along a biogeomorphic gradient. Wilcoxon matched-pairs signed-ranks tests were used to compare the median value of the observed and predicted TOC values of each age-forest category, with their Z-values and probability associated represented in the figure

Modeling the accumulation of AUTOC

The capacity of the 14 parameters representing the forest structure and hydrological connectivity gradients (*Table 1*) as predictors of AUTOC was first tested individually using Spearman correlation tests and controlling for the effect of "observed TOC" on "TOC residuals" (TOC residuals = -0.275 + 0.206 * observed TOC; $F_{1,109}$=18.94; r^2 = 0.14; P < 0.001). Only three variables had a significant correlation with the TOC residuals, namely SD (ρ = 0.30), IVFx (ρ = -0.44) and IVSa (ρ = -0.24) at a P < 0.05. However, a LME model could be significantly fitted to the whole dataset and is summarized in *Table 2*. The model explained 24% of the variability in AUTOC with a combination of three forest structure variables and one hydrological connectivity

Table 2 Result of the best LME model run between autochthonous organic C accumulated in the first 10 cm of soil (response variable) and different forest structure and hydrological connectivity surrogates (*Table 1*, fixed factors) and plot as random effect. Note that the degrees of freedom are only 109 because six outliers were removed from the initial 117 soil samples (39 plots x 3 soil replicates) due to their abnormally high or low TOC values

	Model fit	Pearson's correlation	Explanatory terms	Type of predictor	Relationship sign	*p*-value	Relative contribution
AUTOC (% of OC with an autochthonous origin)	AIC = 269.6525	R^2=0.24 t=5.9271 d.f.=109 *p*<0.001	logIVFx	Forest structure	-	0.0002	39%
			logSD	Forest structure	+	0.0006	27%
			WT	Hydrological connectivity	+	0.0001	24%
			logIVPn	Forest structure	+	0.0036	10%

variable. In particular, it predicted higher AUTOC accumulation with lower relative dominance of *Fraxinus angustifolia*, higher stem densities, higher importance of *Populus nigra* (the three relationships following a logarithmic ratio) and deeper water table levels (i.e., less subsurface hydrological connection). The joint contribution to the model of the forest structure variables was 76%, versus 24% explained by the hydrogeomorphic regime surrogate. Six topsoil samples of the 117 that originated from randomly distributed plots of young and old forest patches were considered to be outliers and therefore removed from the statistical analyses. The analytical results obtained for the samples showed abnormally low or high TOC values detected during the exploration of the residuals of preliminary models and assumed to be incorrectly collected in the field.

DISCUSSION

Organic C that accumulates in the floodplain soil had a predominant allochthonous origin

The results of our study suggest that most of the organic C that accumulated in the forested floodplain soils had an allochthonous origin; presumably from flood-driven sedimentation processes. This occurred even at the more disconnected sites that were only flooded twice a year and where the forest had established for several decades. In particular, ALLOC accumulation was more than six times more important than AUTOC in forests older than 50 years. At the more connected and younger sites, it was impossible to identify any accumulation of organic C in the soil produced by *in situ* vegetation, reflecting that its importance was marginal relative to that deposited by floods. This was a surprising result as far as vertical litterfall in the Ebro (563 g dry matter m^{-2} yr^{-1}) is an important input of organic matter in the floodplain that is within the range of litterfall values reported in other Mediterranean riparian settings and floodplain forests of the warm temperate zone (Gonzalez et al., 2010b). Although the river has dramatically lost its geomorphic dynamics since the 1980's (Cabezas et al., 2009a), sedimentation still plays a much more important role in organic C accumulation than the autochthonous organic matter produced *in situ* by vegetation. Our results provide evidence to support that ALLOC tends to dominate along the floodplain despite flow regulation and embankment, as long as it remains connected with quasi-annual floods (return period of 2.33 yrs. *sensu* Osterkamp and Hedman, 1982). However, this conclusion does not imply that the loss of geomorphic dynamics caused by human impacts on the river system has not had an effect on ALLOC accumulation. In fact, Cabezas et al. (2009b) showed within the same river reach of the Middle Ebro River, that ALLOC floodplain accumulation rates within a floodplain forest patch dating from at least 1957 and in two oxbow lakes with an origin <1946 were higher before completion of dam constructions built for flood risk reduction and river training works (<1963). Dams and river training works did not prevent overbank flooding and thus it could have been expected that ALLOC accumulation within the studied forest patch and the two oxbow lakes did not decrease. River training works implied the cutting off of meanders, partially disconnecting the forest patch and the oxbow lakes from the main channel and thus being less frequently flooded.

Relative role of the abiotic and biotic factors in controlling AUTOC accumulation

According to our results, we suggest that organic C accumulation was controlled mainly by allogenic forces, namely the hydrological regime that ultimately controlled the sedimentation processes. The role of vegetation in organic C accumulation would be more important as a consequence of its physical modulation of hydrogeomorphic fluxes rather than of its own primary productivity, even at the more disconnected and older sites, where late-seral successional stages of vegetation dominated. The presence of non-pioneer trees such as *Fraxinus angustifolia* and *Ulmus minor* at the older sites seems to indicate that vegetation develops autogenically. However, our results suggested that, even then, the soil evolved mainly allogenically. Whether the vegetation and soil succession drivers are uncoupled or the non-pioneer species are much more controlled by the hydrological regime than expected remains an open question that deserves further study. Unlike pioneer riparian tree species, the ecology of non-pioneer floodplain tree species has surprisingly received very little attention up to date.

The capacity of the hydrological regime as driver of organic C accumulation goes beyond its influence on sedimentation processes since surface and sub-surface hydrological connectivity partially controls the accumulation of AUTOC as well. Our LME model showed that a higher accumulation of AUTOC occurs with deeper water tables. Deeper water tables were correlated with lower flood frequency and duration, probably resulting in lower sedimentation and in turn in a decrease of the importance of ALLOC *vs*. AUTOC along the biogeomorphic gradient from the channel to the outer floodplain. However, a further explanation of this relationship may also be the control exerted by the hydrologic regime both on the production of organic C and on its decomposition. Numerous studies have shown that primary production of floodplain forests and forested wetlands is largely controlled by the hydrologic regime (Megonigal, et al. 1997, Burke et al., 1999, Clawson et al., 2001 and many others). In our study area, González et al. (2010b; 2010c; 2012b) showed that the flooding regime explained up to 35% of the litter production variability, directly or indirectly by its effects on the N and P cycles. They also found a weak but positive relationship between deeper water levels and litter production. Therefore, the higher accumulation of AUTOC with deeper water levels in our LME model could be partly due to a higher litter production in those more oxygenated soils. The influence of hydrological connectivity on organic matter decomposition, and hence on the C cycle, has also been proven before (e.g., Baker et al., 2001; Ozalp et al., 2007). Recurrent floods alternating aerobic conditions are supposed to favor organic matter decomposition (Brinson, 1981; Lockaby et al., 1996). It might be the case that deep water tables in our old plots (>50 yr) did not enhance those wetting and drying cycles and hence slowed down decomposition and favored organic C accumulation. No differences in the magnitude of water table fluctuations were observed between plots in the relatively small study area. Therefore, this variable did not contribute to the models and was not shown. However, water table fluctuations at sites with deep water tables are related to less flood submersion episodes than at sites with higher water tables (see WT and FF, *Table 1*). Although it was out of the scope of our study, decomposition must play a key role in the C net accumulation over the long-term in the floodplain soils. With shallower water tables, the young (<25 yrs.) and mature (25-50 yrs.) plots may also be experiencing higher decomposition rates, ultimately reducing C accumulation.

Although the subsurface hydrological connectivity by means of the water table depth below the floodplain surface explained part of the variability in the accumulation of AUTOC, it was the forest structure that most contributed (76%) to the adjusted LME model. The positive effect of stem density on C accumulation was also probably the result of the higher litter production that exhibit floodplain forest patches with higher stem density (Gonzalez et al., 2010b). The inclusion of the IV of *F. angustifolia* and *P. nigra* in the models might respond to a lower productivity and faster decomposability of the organic matter produced by the former and to a higher productivity and lower decomposability for the later (see respective negative and positive contributions in the LME model, *Table 2*), but more studies should be done to confirm this hypothesis.

Limitations of the study

An important limitation of our study may come from the fact that only the first 10 cm of the soil were collected. In certain locations it may have occurred that one single flood event deposited more than 10 cm of sediment, burying the organic C produced *in situ*. In that case, our methodology could have underestimated the AUTOC because it assumed that the topsoil is representative of a likely heterogeneous soil profile (Cooper et al., 1999; Cierjacks et al., 2011). However, this limitation would probably only affect the sites where accumulation is higher than erosion. For example, most of the young plots were covered by gravel, cobbles or sand depositions, and it is very unlikely that a much less denser material such as organic matter has been buried instead of been flushed away. Even at sites where net accretion occurred in our study area, it has been shown that it does at a rate much lower than 10 cm yr^{-1} (Cabezas et al., 2009b) and that the higher rates are localized on narrow margins adjacent to the main channel and on small areas that were the first to be submerged during overbank flooding (Cabezas et al., 2010).

Another source of error may have come from our predictive function ALLOC = f(fine-texture) that was built from one single flood event only and could have been biased towards the ALLOC or AUTOC fraction if its organic matter load was respectively higher or lower than the average of all floods. Different flood events may have different sediment-TOC relationships depending on the flood characteristics in respect to magnitude and flow depths and velocities, timing, frequency, as well as sediment availability and fluxes. Although this is a major limitation of our study, the marginal weight of AUTOC relative to ALLOC (more than six times lower) allows the assumption that with a more consistent predictive function even major deviances from our estimate values would not substantially change the conclusions drawn from our work. For example, the sampled flood in January 2006 did not inundate all 39 plots. Thus, if larger floods inundating all these plots had been sampled and included in the fine sediment fraction-TOC function for ALLOC, the sedimentation pattern would have represented the entire floodplain. Furthermore, this would partially account for the effect of the high spatial variability of sedimentation dynamics and quality (Cabezas et al., 2010). Unfortunately, it is hard to predict the timing of high magnitude, exceptional floods, which makes logistics regarding sample design, i.e. the setup of mats covering the whole natural hydrogeomorphic gradient in the floodplain, very difficult.

Our approach also assumed that the relationship between TOC and the fine sediment fraction was constant during the 50 years covered by the chronosequence. Given that flood control measures have been carried out for decades, sediment sources and availability controlling quantities and quality (e.g. texture, organic matter content) of overbank deposits might have changed over the years. We believe that this assumption, as well as the fact that our model was based on one single flood sampled only, may have influenced our ALLOC/AUTOC ratio quantitatively, but not qualitatively.

Because of the limitations of this exploratory study, the soil texture method applied to identify autochthonous vs. allochthonous C in floodplains (Cabezas and Comín, 2010) needs to be further validated in order to be used more widely within different biogeomorphological settings and according to varying flood characteristics. It would benefit from future comparisons with alternative methods of partitioning AUTOC and ALLOC, such as exploring C budgeting (C depositional fluxes, litterfall fluxes, litterfall decomposition and soil respiration) and C/N ratios, carrying out isotopic characterization or expressing soil C net accumulation rates on a mass basis (g C m^{-2} yr^{-1}) instead of a relative concentration (%).

CONCLUSIONS

This study showed that the accumulation of autochthonous organic C in the Ebro floodplain topsoil only occurred in floodplain forests older than 50 year-old, but even then, it was more than six-fold less abundant than that with an allochthonous origin. Flow regulation and embankment may have substantially reduced the hydrogeomorphic dynamism of the river, but sedimentation by overbank flooding still occurs and probably controls floodplain soil succession and organic C accumulation in the floodplain. Despite the abovementioned limitations of the present study, a promising method to distinguish the relative contributions of autochthonous versus allochthonous C in floodplains was presented here. Thus, this study eventually contributes to the better insights of the role of floodplains in the overall carbon budget and in providing ecosystem services, such as water quality improvement or climate regulation.

Acknowledgements

The authors thank D Jimenez, A de Frutos and MC Sancho for field and lab assistance. This research was funded by the Departments of the Environment (Reserva Natural Galachos) and Science, Technology and University (Research group E-61 on Ecological Restoration)—Aragon Government—and Ministry of Science and Innovation of Spain—MICINN (CGL2008—05153-C02-01/BOS). The first author was granted by the Ministry of Education and Science of Spain—MEC (FPU program).

References

Asselman, N.E.M., Middelkoop, H. 1995. Floodplain sedimentation – quantities, patterns and processes. *Earth Surface Processes and Landforms* 20, 481–499. DOI: 10.1002/esp.3290200602

Baker, III T.T., Lockaby, B.G., Conner, W.H., Meier, C.E., Stanturf, J.A., Burke, M.K. 2001. Leaf litter decomposition and nutrient dynamics in four Southern forested floodplain communities. *Soil Science Society of America Journal* 65, 1334–1347. DOI:10.2136/sssaj2001.6541334x

Bechtold, J.S., Naiman, R.J. 2006. Soil texture and nitrogen mineralization potential across a riparian toposequence in a semi-arid savanna. *Soil, Biology and Biochemistry* 38, 1325-1333. DOI: 10.1016/j.soilbio.2005.09.028

Bendix, J., Hupp, C.R. 2000. Hydrological and geomorphological impacts on riparian plant communities. *Hydrological Processes* 14, 2977–2990. DOI: 10.1002/1099-1085(200011/12) 14:16/17<2977::AID-HYP130>3.0.CO;2-4

Brinson, M. 1981. Primary productivity, decomposition and consumer activity in freshwater wetlands. *Annual Review of Ecology, Evolution and Systematics* 12, 123–161. DOI: 10.1146/annurev.es.12.110181.001011

Burke, M.K., Lockaby, B.G., Conner, W.H. 1999. Aboveground production and nutrient circulation along a flooding gradient in a South Carolina Coastal Plain forest. *Canadian Journal of Forest Research* 29, 1402–1418. DOI: 10.1139/x99-111

Cabezas, A., Comín, F.A., Beguería, S., Trabucchi, M. 2009a. Hydrologic and landscape changes in the Middle Ebro River (NE Spain): implications for restoration and management. *Hydrology and Earth System Sciences* 13, 273–284. DOI:10.5194/hess-13-273-2009

Cabezas, A., Comín, F.A., Walling, D.E. 2009b. Changing patterns of organic carbon and nitrogen accretion on the middle Ebro floodplain (NE Spain). *Ecological Engineering* 35, 1547-1558. DOI: 10.1016/j.ecoleng.2009.07.006

Cabezas, A., Comín, F.A. 2010. Carbon and nitrogen accretion in the topsoil of the Middle Ebro River Floodplains (NE Spain): Implications for their ecological restoration. *Ecological Engineering* 36, 640–652. DOI: 10.1016/j.ecoleng.2008.07.021

Cabezas, A., Angulo-Martínez, M., Gonzalez-Sanchis, M., Jimenez, J.J., Comín, F.A. 2010. Spatial variability in floodplain sedimentation: the use of generalized linear mixed-effects models. *Hydrology and Earth System Sciences* 14, 1655–1668. DOI: 10.5194/hess-14-1655-2010

Chevan, A., Sutherland, M. 1991. Hierarchical partitioning. *The American Statistician* 45, 90–96. DOI: 10.2307/2684366

Cierjacks, A., Kleinschmit, B., Kowarik, I., Graf, M., Lang, F. 2011. Organic matter distribution in floodplains can be predicted using spatial and vegetation structure data. *River Research and Applications* 27, 1048–1057. DOI: 10.1002/rra.1409

Clawson, R.G., Lockaby, B.G., Rummer, B. 2001. Changes in production and nutrient cycling across a wetness gradient within a floodplain forest. *Ecosystems* 4, 126–138. DOI: 10.1007/s100210000063

Clements, F.E. 1916. Plant succession: an analysis of the development of vegetation. Carnegie Institution of Washington, Monograph Series 242. Carnegie Institution, Washington, DC.

Cooper, D.J., Merritt, D.M., Andersen, D.C., Chimner, R.A. 1999. Factors controlling the establishment of Fremont cottonwood seedlings on the Upper Green River, USA. *Regulated Rivers: Research and Management* 15, 419–440. DOI: 10.1002/(SICI)1099-1646(199909/10)15:5<419::AID-RRR555>3.0.CO;2-Y

Corenblit, D., Tabacchi, E., Steiger, J., Gurnell, A.M. 2007. Reciprocal interactions and adjustments between fluvial landforms and vegetation dynamics in river corridors: a review of complementary approaches. *Earth-Science Reviews* 84, 56–86. DOI: 10.1016/j.earscirev.2007.05.004

Crawley, M.J. 2002. Statistical computing. An introduction to data analysis using S-Plus. John Wiley & Sons Ltd., Chichester, England

Daniels, J.M. 2003. Floodplain aggradation and pedogenesis in a semiarid environment. *Geomorphology* 56, 225–242. DOI: 10.1016/s0169-555x(03)00153-3

Day, J.W., Ko, J.Y., Rybczyk, J., Sabins, D., Bean, R., Berthelot, G., Brantley, C., Cardoch, L., Conner, W., Day, J.N., Englande, A.J., Feagley, S., Hyfield, E., Lane, R., Lindsey, J., Mitsch, J., Reyes, E., Twilley, R. 2004. The use of wetlands in the Mississippi Delta for wastewater assimilation: a review. *Ocean and Coastal Management* 47, 671–691. DOI: 10.1016/j.ocecoaman.2004.12.007

Drouin, A., Saint-Laurent, D., Lavoie, L., Ouellet, C. 2011. High-precision elevation model to evaluate the spatial distribution of soil organic carbon in active floodplains. *Wetlands* 31, 1151-1164. DOI: 10.1007/s13157-011-0226-z

Friedman, J.M., Osterkamp, W.R., Lewis, W.M. Jr. 1996. Channel narrowing and vegetation development following a Great Plains flood. *Ecology* 77, 2167–2181. DOI:

Geerling, G.W., Ragas, A.M.J., Leuven, R., van den Berg, J.H., Breedveld, M., Liefhebber, D., Smits, A.J.M. 2006. Succession and rejuvenation in floodplains along the river Allier (France). *Hydrobiologia* 565, 71–86. DOI: 10.2307/2265710

Gergel, S.E., Dixon, M.D., Turner, M.G. 2002. Consequences of human-altered floods: levees, floods, and floodplain forests along the Wisconsin River. *Ecological Applications* 12, 1755–1770. DOI: 10.1890/1051-0761(2002)012[1755:cohafl]2.0.co;2

González, E. 2012a. Seasonal patterns of litterfall in the floodplain forest of a large Mediterranean river. *Limnetica* 31, 173-186.

González, E. 2012b. The ecology of the Middle Ebro floodplain forests and their hydrogeomorphic drivers: An integrative assessment for management. *Méditerranée* 118, 29–40. DOI: 10.4000/mediterranee.6198

González, E., González-Sanchis, M., Cabezas, A., Comín, F.A., Muller, E. 2010a. Recent changes in the riparian forest of a large regulated Mediterranean river: Implications for management. *Environmental Management* 45, 669–681. DOI: 10.1007/s00267-010-9441-2

González, E., Muller, E., Gallardo, B., Comín, F.A., González-Sanchis, M. 2010b. Factors controlling litter production in a large Mediterranean river floodplain forest. *Canadian Journal of Forest Research* 40, 1968–1709. DOI: 10.1139/x10-102

González, E., Muller, E., Comín, F.A., González-Sanchis, M. 2010c. Leaf nutrient concentration as an indicator of Populus and Tamarix response to flooding. *Perspectives in Plant Ecology, Evolution and Systematics* 12, 257–266. DOI: 10.1016/j.ppees.2010.07.001

González, E., González-Sanchis, M., Comín, F.A., Muller, E. 2012. Hydrologic thresholds for riparian forest conservation in a regulated large Mediterranean River. *River Research and Applications* 28, 71–80. DOI: 10.1002/rra.1436

Hughes, F.M.R. 1997. Floodplain biogeomorphology. *Progress in Physical Geography* 21, 501–529. DOI: 10.1177/030913339702100402

Hupp, C.R. 2000. Hydrology, geomorphology and vegetation of Coastal Plain rivers in the South-eastern USA. *Hydrological processes* 14, 2991–3010. DOI: 10.1002/1099-1085(200011/12)14:16/17<2991::aid-hyp131>3.0.co;2-h

IPCC: Working Group III Report "Mitigation of Climate Change", Cambridge University Press, Cambridge, United Kingdom, 2007.

Johnston, C.A. 1991. Sediment and nutrient retention by freshwater wetlands – effects on surface-water quality. *Critical Reviews in Environmental Control* 21, 491–565. DOI: 10.1080/10643389109388425

Lockaby, B.G., Murphy, A.L., Somers, G.L. 1996. Hydroperiod influences on nutrient dynamics in decomposing litter of a floodplain forest. *Soil Science Society of America Journal* 60, 1267–1272. DOI: 10.2136/sssaj1996.03615995006000040044x

Magdaleno, F., Fernandez, J.A. 2011. Hydromorphological alteration of a large Mediterranean river: relative role of high and low flows on the evolution of riparian forests and channel morphology. *River Research and Applications* 27, 374–387. DOI: 10.1002/rra.1368

Megonigal, J.P., Conner, W.H., Kroeger, S., Sharitz, R.R. 1997. Aboveground production in Southeastern floodplain forests: A test of the subsidy-stress hypothesis. *Ecology* 78, 370–384. DOI: 10.2307/2266014

Naiman, R.J., Décamps, H., McClain, M.E. 2005. Riparia. Ecology, conservation, and management of streamside communities. Elsevier Academic Press. San Diego

Noe, G.B., Hupp, C.R. 2005. Carbon, nitrogen, and phosphorus accumulation in floodplains of Atlantic Coastal Plain rivers USA. *Ecological Applications* 15, 1178–1190. DOI: 10.1890/04-1677

Ollero, A. 1995. Dinámica reciente del cauce del Ebro en la Reserva Natural de los Galachos (Zaragoza). *Cuaternario y Geomorfología* 9, 85–93.

Ollero, A. 2007. Channel adjustments, floodplain changes and riparian ecosystems of the Middle Ebro River: assessment and management. *Water Resources Development* 23, 73–90. DOI: 10.1080/07900620601159586

Osterkamp, W.R., Hedman, E.R. 1982. Perennial-streamflow Characteristics Related to Channel Geometry and Sediment in Missouri River Basin. U.S. Geological Survey Professional Paper, vol. 1242. U.S. Geological Survey, Washington, DC.

Ozalp, M., Conner, W.H., Lockaby, B.G. 2007. Above-ground productivity and litter decomposition in a tidal freshwater forested wetland on Bull Island, SC, USA. *Forest Ecology and Management* 245, 31–43. DOI: 10.1016/j.foreco.2007.03.063

Piégay, H., Hupp, C.R., Citterio, A., Moulin, B., Walling, D.E. 2008. Spatial and temporal variability in sedimentation rates associated with cutoff channel infill deposits: Ain River, France. *Water Resources Research* 44, W05420, DOI: 10.1029/2006WR005260.

Pinay, G., Fabre, A., Vervier, Ph., Gazelle, F. 1992. Control of C, N, P distribution in soils of riparian forests. *Landscape Ecology* 6, 121–132. DOI: 10.1007/bf00130025

Pinheiro, J.C., Bates, D.M. 2000. Mixed-effects models in S and S-PLUS. Springer, New York, USA

Pinheiro, J.C., Bates, D.M., DebRoy, S., Sarkar, D., R Development Core Team. 2007. nlme: Linear and nonlinear mixed effects models. R package version 3.1–86.

R Development Core Team (2011). R: A language and environment for statistical computing. R Foundation for Statistical Computing, Vienna, Austria. ISBN 3-900051-07-0, URL http://www.R-project.org/

Steiger, J., Gurnell, A.M., Ergenzinger, P., Snelder, D. 2001. Sedimentation in the riparian zone of an incising river. *Earth Surface Processes and Landforms* 26, 91–108. DOI: 10.1002/1096-9837(200101)26:1<91::aid-esp164>3.0.co;2-u

Steiger, J., Gurnell, A.M. 2003. Spatial hydrogeomorphological influences on sediment and nutrient deposition in riparian zones: observations from the Garonne River, France. *Geomorphology* 49, 1–23. DOI: 10.1016/s0169-555x(02)00144-7

Steiger, J., Gurnell, A.M., Goodson J. 2003. Quantifying and characterizing contemporary riparian sedimentation. *River Research and Applications* 19, 335–352. DOI: 10.1002/rra.708

Verhoeven, J.T.A., Arheimer, B., Yin, C.Q., Hefting, M.M. 2006. Regional and global concerns over wetlands and water quality. *Trends in Ecology and Evolution* 21, 96–103. DOI: 10.1016/j.tree.2005.11.015

Walling, D.E., He, Q. 1997. Investigating spatial patterns of overbank sedimentation on river floodplains. *Water, Air and Soil Pollution* 99, 9–20. DOI: 10.1007/bf02406840

Walsh, C., McNally, R. 2007. hier.part: Hierarchical Partitioning. R package version 1.0–2. Available from www.r-project.org.

Wigginton, J.D., Lockaby, B.G., Trettin, C.C. 2000. Soil organic matter formation and sequestration across a forested floodplain chronosequence. *Ecological Engineering* 15, S141–S155. DOI: 10.1016/s0925-8574(99)00080-4

JOURNAL OF ENVIRONMENTAL GEOGRAPHY
Journal of Environmental Geography 7 (3–4), 39–47.
DOI: 10.2478/jengeo-2014-0005
ISSN: 2060-467X

ENVIRONMENTAL CHANGES IN HISTORICAL TIMES NEAR APOSTAG ON THE DANUBE-TISZA INTERFLUVE, HUNGARY (A COMPLEX RESEARCH BASED ON ARCHAEOLOGICAL EXCAVATION AND GEOMORPHOLOGICAL INVESTIGATIONS)

Diána Nyári[1*], István Knipl[2], György Sipos[1], Tímea Kiss[1]

[1]Department of Physical Geography and Geoinformatics, University of Szeged, Egyetem u. 2-6, H-6722 Szeged, Hungary
[2] Museum of Kiskunfélegyháza, Dr. Holló Lajos u. 9, H-6100 Kiskunfélegyháza, Humgary
*Corresponding author, e-mail: nyaridia@gmail.com

Research article, received 26 February 2014, accepted 31 March 2014

Abstract

The sensitive, partly fixed dune areas are good indicators of alteration, since they react rapidly to changing environmental conditions. Due to the climate changes, especially the increased aridity during the Holocene, many blown sand areas became active. Later, humanity had increasing impact of on its environment, thus sand movements occurred due to anthropogenic activities. Aeolian activities were identified not only in the historical times but also a few decades ago, when the moving sand caused significant problems on surfaces becoming bare. The present work will provide good evidence on sand movement in historical times caused by human impact on the environment with the help of OSL dating and archaeological research in the vicinity of the town of Apostag, which is located on the largest blown-sand area of Hungary on the Danube-Tisza Interfluve. The aims of the research were to identify the ethnical groups and their possible activities; to map the geomorphology of the study area; to determine the periods of aeolian activity; to assign the possible types of human activities in connection with climatic changes enabling aeolian activity.

Keywords: environmental changes, optically stimulated luminescence, archaeology, geomorphology, blown sand movement

INTRODUCTION

The sensitive, partly fixed dune areas are good indicators of alteration, since they react rapidly to changing environmental conditions. It can be confirmed by the result of an Australian survey where sand sheet was detected in a fixed dune area covered by forests due to sand movement in the early Holocene. The palinological investigations did not confirm aridification in the area; the precipitation increased continuously after the Pleistocene and reached the maximum 4000 years ago (Shulmeister, 1992; Shulmeister and Lees, 1992). However, the presence of the dry period is confirmed by many proofs; the change was probably so rapid and the dry period lasted so shortly that its influence could not be detected in the pollen spectrum (Nott et al., 1999). Therefore, the investigation of sand movement has an important role in the assessment of environmental impacts, since small changes can modify the sensitive balance and can indicate mobilisation.

Due to the climate changes, especially the increased aridity during the Holocene, many blown sand areas became active e.g. in the United States (Forman et al., 1992, 1995; Olson et al., 1997; Rawling et al., 2003; Stokes and Swinehart, 1997; Wilkins and Currey, 1999). On the European coastal areas climate change due to the Little Ice Age indicated sand movements (Borja et al., 1999; Clarke et al., 2002, Wilson and Braley, 1997; Wintle et al., 1998). Forest fires also resulted in bare surfaces where wind erosion could become dominant (Filion, 1984; Filion et al., 1991; Kayhkö et al., 1999). Later, humanity had increasing impact of on its environment, thus sand movements occurred due to anthropogenic activities (Wilkins and Currey, 1999; Wilson and Braley, 1997; Wintle et al., 1998). Thus, the investigation of Holocene sand movements has an important role in the assessment of global warming, climate change and the human impact on the environment.

Sand movements were identified not only in the historical times but also a few decades ago, when the moving sand caused significant problems on surfaces becoming bare. In Ireland the improper land use in the 1930-40s caused the degradation of cultivated areas, and the huge amount of sand blocked roads (Wilson and Braley, 1997). Increased aeolian activities were identified in Canada in the 1920-30s (Lemmen et al., 1998), furthermore during the 1980s as well (David and Wolfe, 1997). Huge damages were also reported in the USA in the beginning of 1930s due to wind erosion (Orlove, 2005). Sand movements caused significant damages in the lack of vegetation on the surface also in Hungary (Mezősi and Szatmári, 1998; Szatmári, 2004). According to Borsy (1972) 5-40 cm thick sand layers was transported during a few days south-east from Kiskunhalas in April 1967. Significant damages due to deflation on sandy surfaces were recorded from Nyírség region between 10 and 12 February 1984 (Lóki, 1985).

Climate change, growing population, the development of agricultural techniques and the changes in land use caused human induced environmental changes, which became increasingly significant in history. Good examples can be found in Hungary on the Danube-Tisza Interfluve where both the change in climatic conditions and the anthropogenic disturbance caused aeolian activity during historical times. Therefore, the geomorphology of the area transformed and the Pleistocene forms were reshaped by Holocene sand-movements.

The earliest blown sand movements on the Danube-Tisza Interfluve took place in the Inter Pleniglacial of the Pleistocene (Sümegi and Lóki, 1990; Sümegi, 2005) and subsequently there was aeolian activity during the Middle Pleniglacial of the Pleistocene after 25 200 ± 300 year ago (Krolopp et al., 1995; Sümegi, 2005). According to earlier researches on the Danube-Tisza Interfluve the most significant aeolian activity occurred during the Upper Pleniglacial (Borsy, 1977ab, 1987, 1989, 1991; Sümegi et al., 1992; Sümegi and Lóki, 1990; Sümegi, 2005). Later, the two cold and dry periods, the Older Dryas and Younger Dryas in the Pleistocene were convenient for aeolian rework (Borsy et al., 1991; Hertelendi et al., 1993) which is supported by radiometric, optical and thermo-luminescenece measurements too (Gábris et al., 2000, 2002; Gábris, 2003; Újházy, 2002; Újházy et al., 2003).

Sand dunes, formed under cold and dry climate in the Pleistocene, were gradually fixed as the climate changed to warm and humid during the Holocene. However, researchers draw attention to the possibility of sand movement in the Holocene too. The warmest and driest Holocene phase (Boreal Phase) was the most adequate for dune formation (Járainé, 1966, 1969; Borsy, 1977ab, 1987, 1991; Gábris, 2003; Kádár, 1956; Marosi, 1967; Újházy et al., 2003), though, certain investigations claim that the second half of the Atlantic Phase could also be dry enough for the remobilisation of sand (Járainé, 1966, 1969; Borsy and Borsy, 1955; Borsy, 1977ab; Gábris, 2003; Újházy et al., 2003). Nevertheless, the latest, usually local signs of aeolian activity can be related to various types of human impact. Former investigations consider that sand movement could occur during the Turkish occupation (16^{th} - 17^{th} century AD) and subsequently in the 18^{th} -19^{th} century AD due to deforestation (Borsy, 1977ab, 1987, 1991; Marosi, 1967).

Based on archaeological investigations and OSL measurements on the Danube-Tisza Interfluve aeolian activity occured in the Bronze Age (Gábris, 2003; Újházy et al., 2003; Nyári and Kiss, 2005a, b; Kiss et al., 2006; Nyári et al., 2006, 2007a and b; Sipos et al., 2006), then the surface became stable for a long period, until the 3^{rd}-4^{th} centuries AD. As later the climate turned dry (Rácz, 2006; Persaits et al., 2008) and the anthropogenic disturbance became more significant conditions became suitable for aeolian activity, which is proved by several researchers (Lóki and Schweitzer, 2001; Kiss et al., 2006; Nyári et al., 2006, 2007a, b; Sipos et al., 2006). Sand movement was also characteristic in the Migration Period, especially during the 6^{th}-8^{th} century AD, which was the realm of the Avars (Nyári and Kiss, 2005a,b; Kiss et al., 2006; Nyári et al., 2006, 2007a, b; Sipos et al., 2006) Subsequent aeolian activity occurred also in the Árpád Age (11^{th}-13^{th} c. AD, Lóki and Schweitzer, 2001; Gábris, 2003; Újházy et al., 2003; Nyári et al., 2006) and when the Cumanians inhabited the territory (13^{th} c. AD, Sümegi, 2001; Kiss et al., 2006; Nyári et al., 2006, 2007a, b; Sipos et al., 2006). The latest aeolian activity occurred in the 15^{th} century AD (Nyári et al., 2007a).

The present work will provide good evidence on sand movement in historical times caused by human impact on the environment with the help of OSL dating and archaeological research in the vicinity of the town of Apostag, which is located on the largest blown-sand area of Hungary on the Danube-Tisza Interfluve. The aims of the research were to identify the ethnical groups and their possible activities; to map the geomorphology of the study area; to determine the periods of aeolian activity; to assign the possible types of human activities in connection with climatic changes enabling aeolian activity.

STUDY AREA

The study area is located on the west part of the Danube-Tisza alluvial fan, south from Apostag *(Fig. 1.)*. The archaeological sites are situated on the south part of the study area *(Fig. 8.)*.

METHODS

Archaeological investigations

The archaeological investigations enabled to determine the historical characteristic of the study area and to assign the activity of the seated people who inhabited the area.

Geomorphological mapping

The relief and geomorphological map of the investigated area was compiled on the basis of field measurements and 1:10,000 scale topographic maps. Forms typical on stabilised blown-sand areas were allocated — blowout depressions, blowout ridges, hummocks, and the brink lines of dunes — and also former flows could be determined.

OSL measurements

Four samples were collected from two profiles. Extraction and sample preparation procedures followed the steps introduced by Aitken (1998) and Mauz (2002) and aimed at the separation of quartz grains of suitable (90-150 μm) size. Measurements were made on an automated RISOE TL/OSL-DA-15 type luminescence reader. Throughout the measurements the SAR technique, described in detail by Murray and Wintle (2000), was followed.

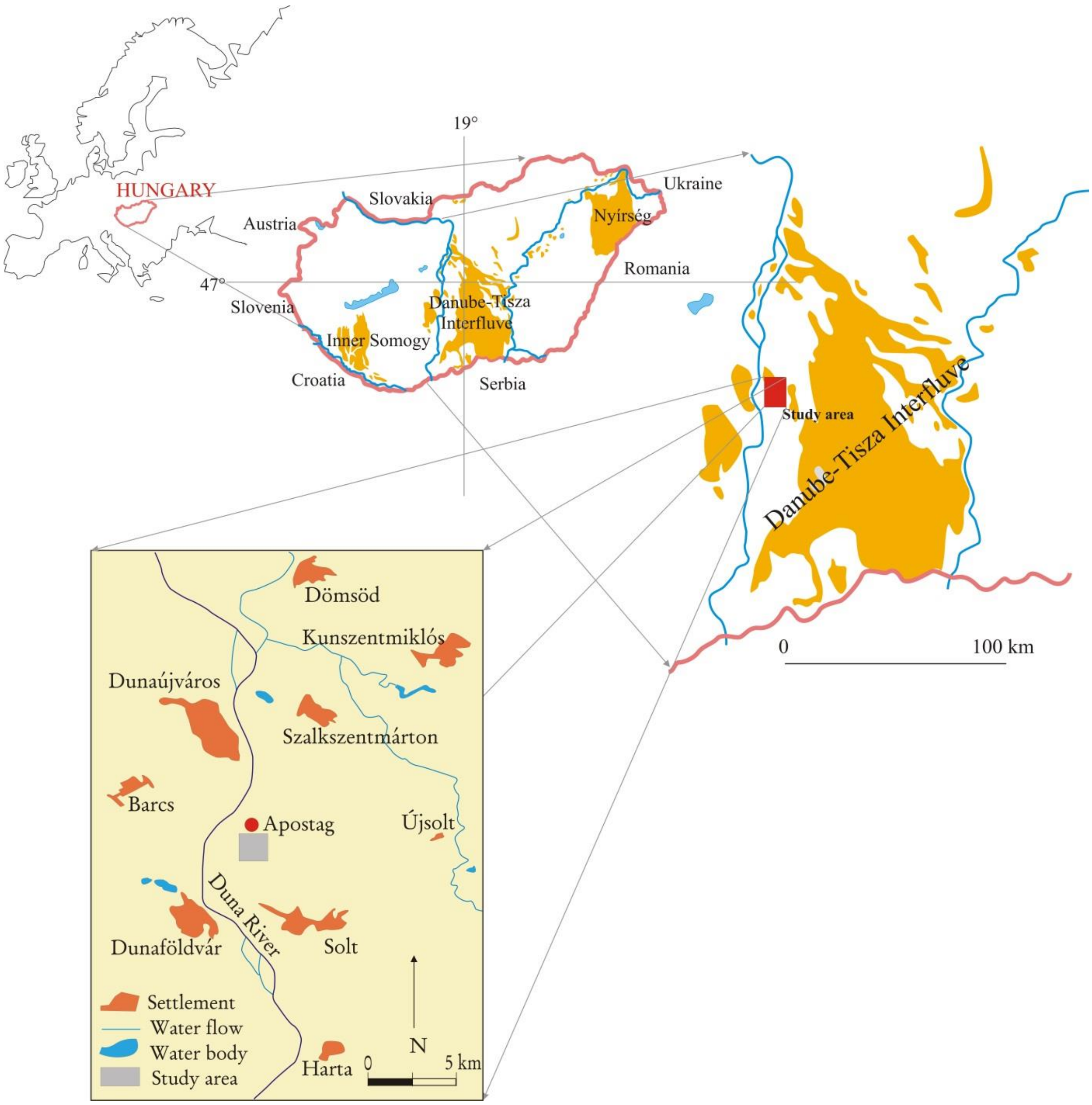

Fig. 1 Location of the study area

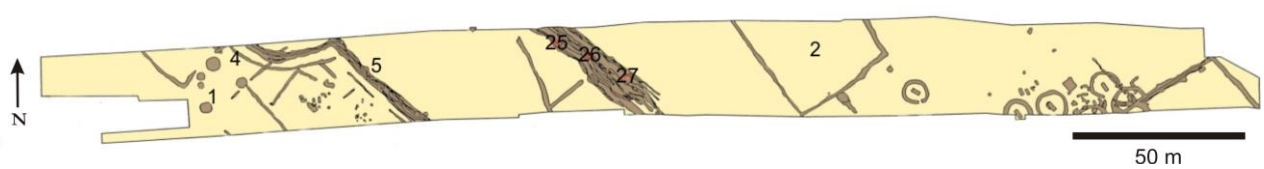

Fig. 2 The archaeological site and the excavated features
1: Pit, 2: Stock yard, 3-5, 25-27: Ditches

RESULTS

Archaeological research

During rescue excavation on the area about 300x25 meters 34 chases, 38 holes, 15 pile holes, 3 pits, nine graves from the sarmatian era as well as a modern pit were explored. The objects of the settlement were sometimes dug on each other, and often they formed small groups and were situated quite rarely. The excavated area was arranged into well segregated parts. On the western part next to an expansive ditch system, on stock-raising referring hurdle ditches and pits showed up, in the middle another large ditch system, and on the eastern part scarce marks of settlement and a few graves of the cemetery were founded (*Fig. 2*). On the western edge of the rescue excavated area, close to the animal keeping hurdles three middle-sized pits were excavated, which included no rateable appendices; only in one of the pits a skeleton of a dog was found (*Fig. 3*). In addition on the middle of the area a stock-yard was found (*Fig. 2*).

Fig. 3 Skeleton of a dog

In the examined area various ditches could be observed. Firstly there were extensive hurdle ditches connected to the animal keeping part of the settlement; furthermore, large sections of ditch systems could be observed, which were probably used for some kind of defence of the settlement. The most western parts of one of the ditch systems were the ditches nr. 5. and 4. The narrow, shallow ditch nr.4 followed the line of the ditch nr. 5. (*Fig. 2*). The 2 m wide, variable deep, arranged bottomed 5. ditch cut the excavated area in northwest- southeast direction across. Its cutting was extremely interesting: in almost its whole length in both sides of the ditch there was a sinking: it was 20-30 cm wide and came down with the same depth under the bottom level of the ditch, which is supposed to be the basic ditch of the pile lines on both sides of the ditch (Fig.4). The other great, with the previous parallel ditch system formed by the ditches 25-26-27, turned up about 40-45 m from the ditches 4-5 (*Fig. 2*). The depth of the three parallel and approaching ditches were remarkably distinct.

Fig. 4 Ditch Nr. 5

The middle and the eastern (26-27.) ditches had arranged bottoms, but according to the pruned surface they were shallow, however the ditch nr. 25 ended about 80 cm from the pruned level. Its width was about 200 cm; its lateral wall was upright, partly splay, and its bottom was mainly straight. The speciality of this ditch system is that with the sudden “jumping up” of the ditch bottom it became really shallow on a 15 m long stage. While there is no mark on a subsequent infilling of the ditch, we recognise the shallow part as the part of the original ditch. The explanation can be that there might be some kind of entrance or passage through the ditch (*Fig. 5*). According to the air photo made on the area the ditch system can be followed on few hundred meters. In the long run we can assume only generally, that it might be a part of a significant fortification-ditch defence system (*Fig. 6*).

Fig. 5 Ditch Nr. 25

The air photos show not only the great fortification system but also a wide cemetery (*Fig. 6*). On the eastern end of the excavated area nine graves were rescue excavated, with and without round-ditch (*Fig. 2.*). The NW-SE sited graves were dominantly close to each other. It is remarkable that graves without round-ditches were immediately next to those with round-ditch, close to each other, in a row. Usually young deceased were buried in those graves, while in the round-ditched graves rather adults (*Fig. 7 a, b*). Most of these were robbed.

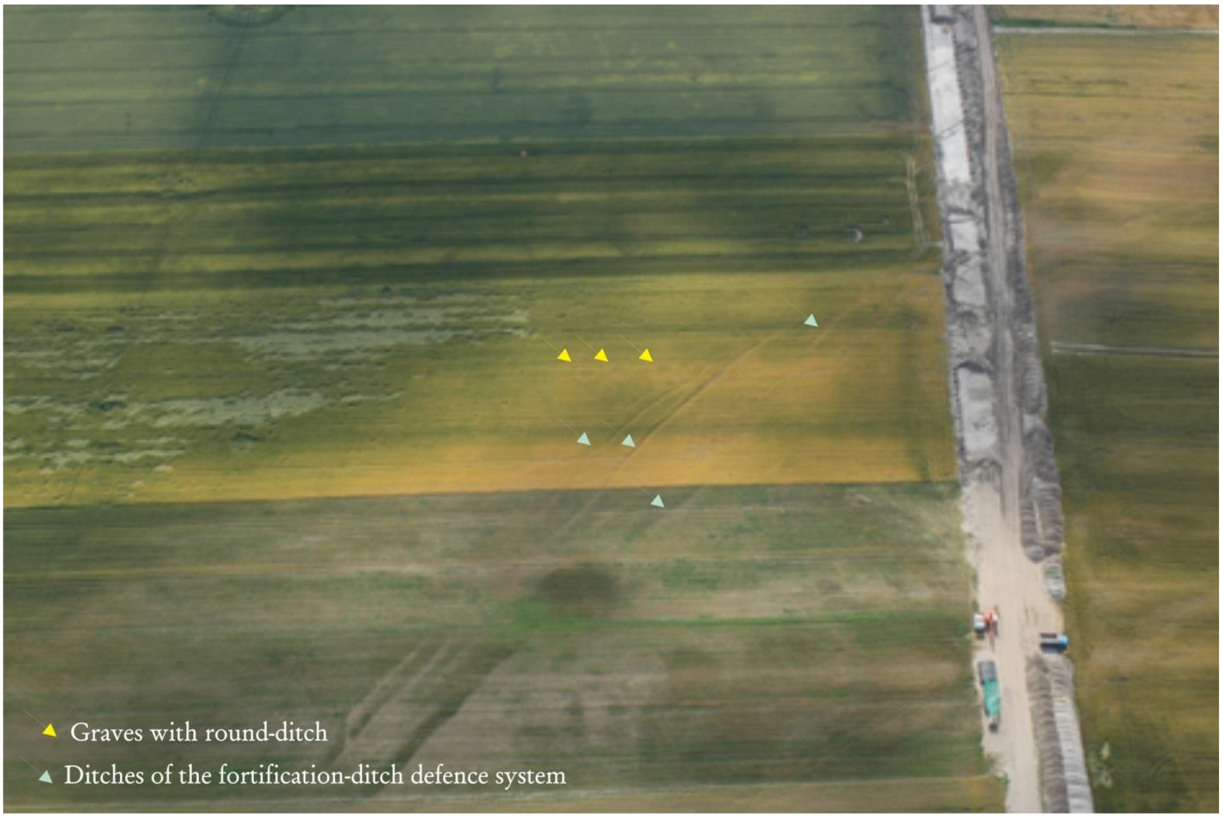

Fig.6 Air photo of the excavated area and its neighbourhood

The artefacts of untouched graves allowed of an interesting observation on costume history. In two graves a significant amount of different coloured pastry beads were found *(Fig. 7c)*. The beads were in rows, in the front part of the neck and on the wrist, so they might be the decoration of the neck and sleeve of the clothes. In the graves some brazen and silver fibulas, mostly close to the breastbone, were found and next to one skeleton there were two fibulas under each other. Other grave artefacts are: two brazen earrings, one brazen ring, one spindle knob, fragments of two small pots, fragments of an iron knife and iron sword, together with a coin from Roman age (Aurelianus: 270-275).

Geomorphological mapping

The mapped area is 4 km^2 and situated on the western part of the Danube-Tisza Interfluve 2.5 kilometers from the Danube river on the border line between a stabilized blown sand surface and a former floodplain of the Danube River (*Fig. 1*). The altitude of the area varies between 94 and 102 m asl. In the middle a sand deposit can be seen which has been blown there by the wind from the deposited sand of the Danube River. This higher sandy place is bordered by low lying, flat areas (*Fig. 8*). Based on the relief map the centre of the investigated area represents an accumulation zone,

a. b. c.

Fig. 7 a: robbed adult grave; b: a round ditch; c: untouched grave with artefacts

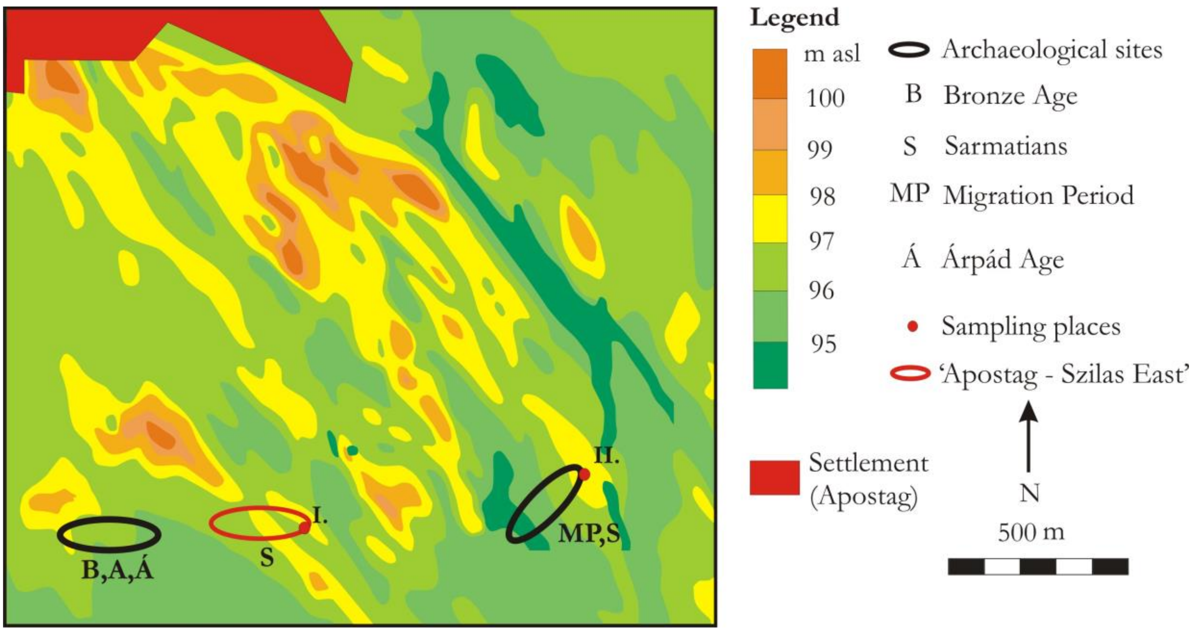

Fig. 8 The relief of the study area, the archaeological sites and the sampling places

where according to the geomorphological mapping the most typical forms are blowout depressions, blowout ridges and blowout dunes (hummocks). Around this higher sandy surface low lying, flat, former alluvial areas of the Danube River can be identified (*Fig. 9*). The Holocene morphological evolution of the investigated area is complex. In most of the cases Pleistocene forms were reshaped and transformed, thus at certain locations the original morphology can hardly be identified. Remobilisation and reshaping was especially intensive during historical times, however it was restricted to smaller patches of land.

Depositional history

Samples were collected at two locations (*Fig. 8*). Profiles show the types of different depositions and OSL ages (*Fig. 10*). Based on the results, OSL yielded Early Holocene age for the lowermost layer (9094 ± 1096 BP), on which sequences of fluvial deposits, paleosoils and blown-sand layers were formed during the Holocene. Initially fluvial processes were characteristic on the territory.

The Danube River deposited carbonate silt on the surface. When the Danube left this area and drifted to West a thick paleosoil was formed on the

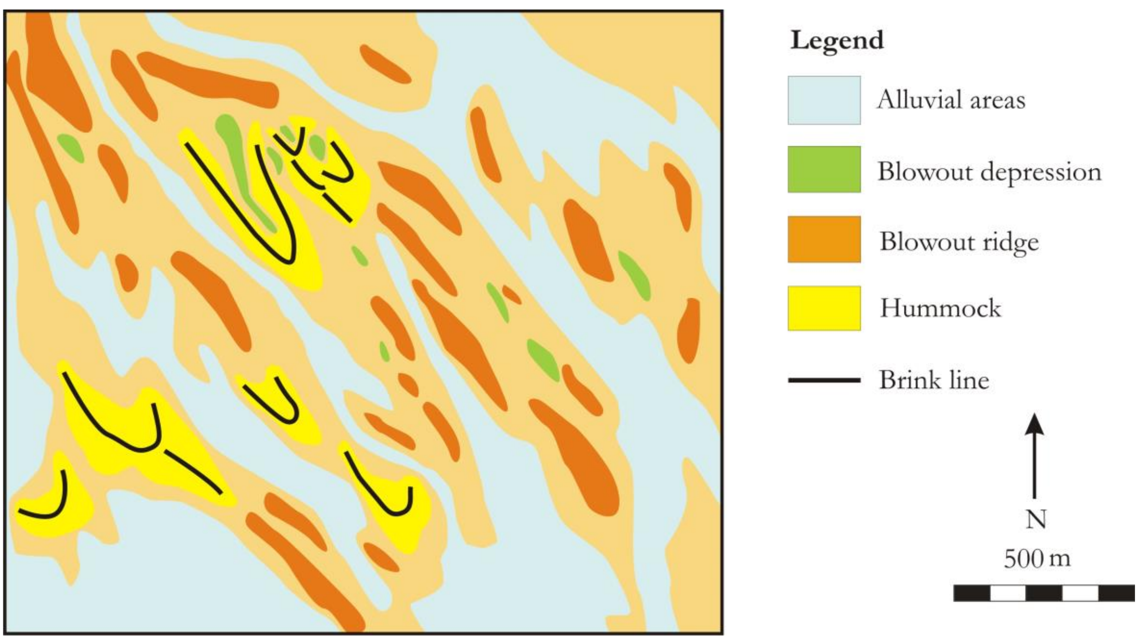

Fig. 9 The geomorphological setting of the study area

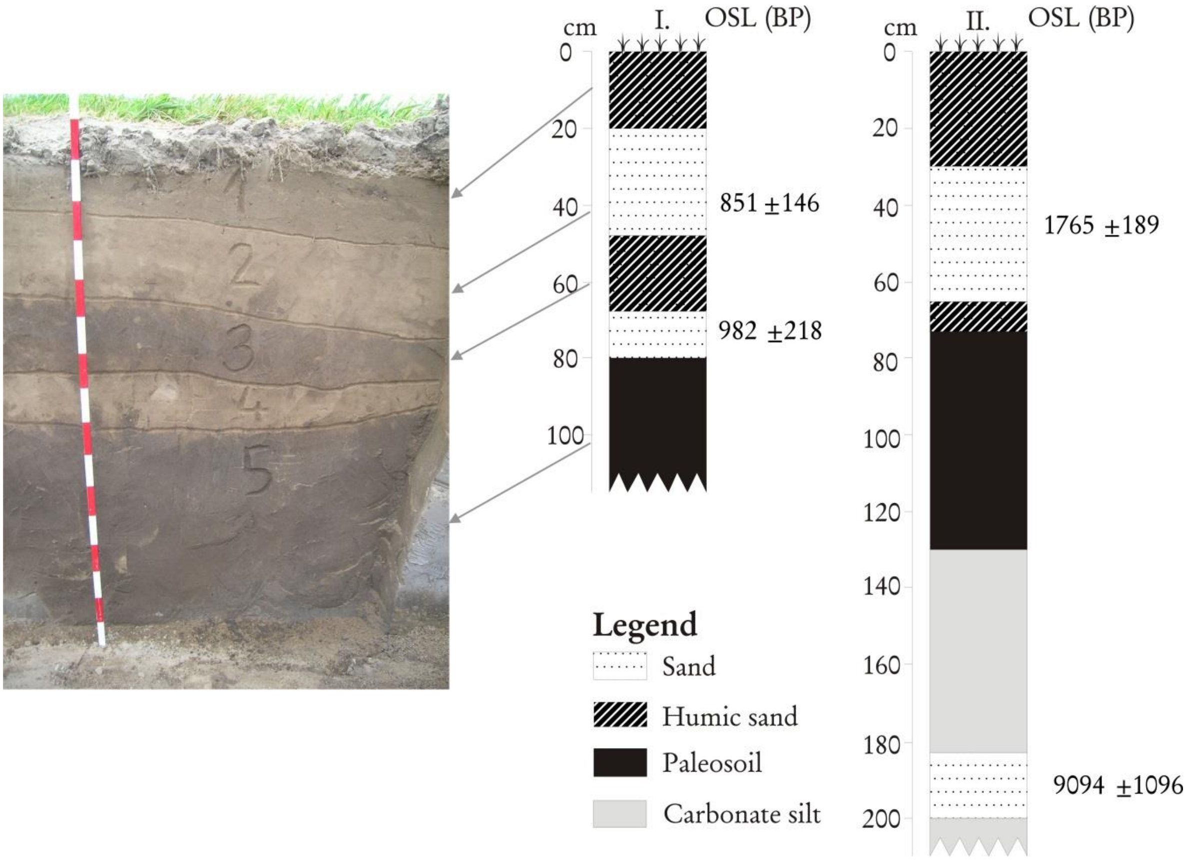

Fig. 10 Profiles of the sampling places, depositions and OSL data

surface. Paleosoil is superimposed by blown-sand layers of varying thickness in the profiles The first phase of sedimentation occurred 1765±189 y BP. Later aeolian activity restarted two times again, first 982 ± 218 y BP then 851 ± 146 y BP, according to OSL measurements

CONCLUSION

Age and sedimentological data of the profiles were compared to archaeological evidence, the spatial distribution of findings of the site (Wicker 2005), as well as archaeological relicts from the area of Apostag (KJM 1968, 1976, 1985, 2001, 2005). This enabled the reconstruction of the type, intensity and the result of human impact on the paleo-environment. All age data were plotted on a historical timescale (*Fig. 11*) According to the archaeological evidences, people settled down on the paleosoil surface. They were Sarmatians who inhabited the area between the 1st and 4th century. They were farmers and they also kept livestock on the pastures. The excavated marks of trenches and stock-yards prove that the excavated site probably functioned as a stock farm and the neighbouring mounds have been pastures or meadows. This is confirmed by the OSL measurements, as blown-sand movement was detected on the nearby higher places in the 3rd century AD (OSL: 1765 ± 189 BP). Probably the cause was ploughing or over-grazing resulting bare surfaces, which were scenes of wind erosion. Finally, a 60-70 cm sand sheet covered the paleosoils of neighbouring mound.

On the evidence of the archaeological investigations, later, in the Árpád Age a larger population lived on the territory of the study area and their activity meant an intensive burden on the environment. Be-

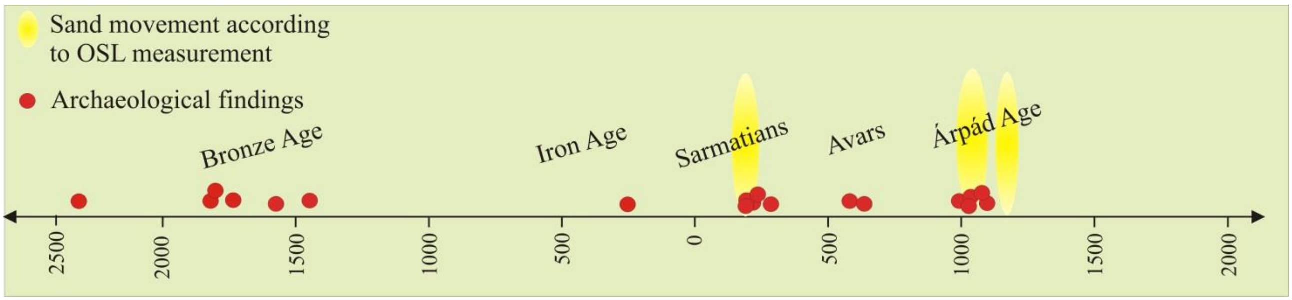

Fig. 11 OSL ages and the archaeological relicts of the area of Apostag

cause of the human impact, aeolian activity revealed again in the 11th century AD (OSL: 982 ± 218 BP) and a 20-30 cm thick sand sheet covered this time the former surface of the excavated site then a poorly developed soil was formed. Afterwards during the 12th century (OSL: 851 ± 146 BP) blown sand movement happened over again and another 60-80 cm thick sand layer covered the territory of the excavated area.

As a conclusion, there was three times spatially localized blown-sand movement on the study area. The first movement effected a sand deposition on the next mound from the archaeological site because of the activity of Sarmatians then two times blown sand movements covered the area of the site by sand sheets as a result of anthropogenic disturbance in the Árpád Age. Thus, former landscape has been changed. Today the surface is approximately 1 m higher than before and a sandy surface can be found where a thick paleosoil was situated before.

Acknowledgements

This research was supported by the European Union and the State of Hungary, co-financed by the European Social Fund in the framework of TÁMOP 4.2.4. A/2-11-1-2012-0001 'National Excellence Program'.

References

Aitken M.J. 1998. An introduction to optical dating: the dating of Quaternary sediments by the use of photon-stimulated luminescence. Oxford: Oxford University Press

Borja, F., Zazo, C., Dabrio, C.J., Díaz del Olmo, F., Goy, J.L., Lario, J. 1999. Holocene Aeolian phases and human settlements along the Atlantic coast of southern Spain. *The Holocene* 9 (3), 333–339. DOI: 10.1191/095968399668924476

Borsi, Z-né, Borsy, Z. 1955. Pollenanalitikai vizsgálatok a Nyírség északi részében. *Közlemények a KLTE Földrajzi Intézetéből* 22, 1–10.

Borsy Z. 1977a. A Duna-Tisza köze homokformái és a homokmozgás szakaszai. *Alföldi tanulmányok* 1, 43–53.

Borsy, Z. 1977b. A magyarországi futóhomok területek felszínfejlıdése. *Földrajzi Közlemények,* 12–16.

Borsy, Z. 1987. Az Alföld hordalékkúpjainak fejlődéstörténete. Nyíregyházi Főiskola Füzetek, 5-37.

Borsy, Z. 1989. Az Alföld hordalékkúpjainak negyedidőszaki fejlődéstörténete. *Földrajzi Közlemények* 211–222.

Borsy, Z. 1991. Blown sand territories in Hungary. *Zeitschrift für Geomorphologie N.F*. Suppl. 90, 1–14.

Borsy, Z., Félegyházi, E., Hertelendi, E., Lóki, J., Sümegi, P. 1991. A bócsai fúrás rétegsorának szedimentológiai, pollenanalitikai és malakofaunisztikai vizsgálata. *Acta Geographica Debrecenina, Tomus* 28–29, 263–277.

Clarke, M., Rendell, H. Tastet, J-P., Clavé, B., Massé, L. 2002. Late-Holocene sand invasion and North Atlantic storminess along the Aquitaine Coast, southwest France. *The Holocene* 12 (2), 231–238. DOI: 10.1191/0959683602hl539rr

Filion, L. 1984. A relationship between dunes, fire and climate recorded in the Holocene deposits of Quebec. *Nature* 309, 543–546. DOI: 10.1038/309543a0

Filion, L., Saint-Laurent, D., Desponts, M., Payette, S. 1991. The late Holocene record of Aeolian and fire activity in northern Quebec, Canada. *The Holocene* 1, 201–208. DOI: 10.1177/095968369100100302

Forman, S.L., Goetz, A.F.H, Yuhas, R.H 1992. Large-scale stabilized dunes on the High Plains of Colorado: understanding the landscape response to Holocene climates with the aid of images from space. *Geology* 20, 145–148. DOI: 10.1130/0091-7613(1992)020<0145:lssdot>2.3.co;2

Forman, S.L., Oglesby, R., Markgraf, V., Stafford, T. 1995. Paleoclimatic significance of Late Quaternary Aeolian deposition on the Peiedmont and High Plains, central United States. *Global and Planetary Change* 11, 35–55. DOI: 10.1016/0921-8181(94)00015-6

Gábris, Gy., Horváth, E., Novothny, Á., Ujházy, K. 2000. Environmental changes during the Last -, Late- and Postglacial in Hungary. In: Kertész, Á., Schweitzer, F. (eds) Physico-geographical Research in Hungary, Studies in Geography in Hungary, 32. Akadémiai Kiadó, Budapest, 47–61.

Gábris, Gy., Horváth, E., Novothny, Á., Ujházy, K. 2002. History of environmental changes from the Glacial period in Hungary. *Prehistoria* 3, 9–22.

Gábris, Gy. 2003. A földtörténet utolsó 30 ezer évének szakaszai és a futóhomok mozgásának főbb periódusai Magyarországon. *Földrajzi Közlemények*, 1–13.

Hertelendi, E., Lóki, J., Sümegi, P. 1993. A Háy-tanya melletti feltárás rétegsorának szedimentológiai és sztatigráfiai elemzése. *Acta Geographica Debrecina* 30-31, 65–75.

Járainé Komlódi, M. 1966. Adatok az Alföld negyedkori klíma és vegetációtörténetéhez. I. Bot. Közlem. 53. 191–200.

Járainé Komlódi, M. 1969. Adatok az Alföld negyedkori klíma és vegetációtörténetéhez. II. Bot. Közlem. 56. 43–55.

Käyhkö, J.A.,Woesley, P., Pye, K., Clarke, M.L. 1999. A revised chronology for Aeolian activity in subarctic Fennoscandia during the Holocene. *The Holocene* 9 (2), 195–205. DOI:

Kádár L. 1956. A magyarországi futóhomok-kutatás eredményei és vitás kérdései. *Földrajzi Közlemények* 4, 143–163.

KJM 1968. Archeological documentation Nr. 134. Katona József Museum, Archeological Database.

KJM 1976. Archeological documentation Nr. 1016. Katona József Museum, Archeological Database.

KJM 1985. Archeological documentation Nr. 87. 519. Katona József Museum, Archeological Database.

KJM 2001. Archeological documentation Nr. 997, 998, 999, 1039, 1040. Katona József Museum, Archeological Database.

KJM 2005. Archeological documentation Nr. 1418, 1419, 1420, 1423, 1463, 1464, 1465. Katona József Museum, Archeological Database.

Kiss, T., Nyári, D., Sipos, Gy. 2006. Blown sand movement in historical times in the territory of Csengele. In: Kiss, A., Mezősi, G., Sümeghy, Z. (eds.) Landscape, Environment and Society. Szeged, 373–383.

Kiss, T., Nyári, D., Sipos, Gy. 2008. Történelmi idők eolikus tevékenységének vizsgálata: A Nyírség és a Duna- Tisza köze összehasonlító elemzése. In Szabó, J., Demeter, G. (eds.) Tanulmányok a Kádár László 100. évfordulóján rendezett tudományos konferenciára. Kossuth Egyetemi Kiadó, Debrecen, 99–106.

Krolopp, E., Sümegi, P. Kuti, L., Hertelendi, E., Kordos, L. 1995. A Szeged-Öthalom környéki löszképződmények keletkezésének paleoökológiai rekonstrukciója. *Földtani Közlemények* 125, 309–361.

Lemmen, D.S., Vance, R.E., Campbell, I.A., David, P.P., Pennock, D.J., Sauchyn, D.J., Wolfe S.A. 1998. Geomorphic systems of the palliser triangle, southern canadian preries: description and response to changing climate. *Geological Survey of Canada, Bulletin* 521, 30–31. DOI: 10.4095/210076

Lóki, J., Schweitzer, F. 2001. Fiatal homokmozgások kormeghatározási kérdései a Duna-Tisza közi régészeti feltárások tükrében. *Papers from the Institute of Geography, University of Debrecen* 221, 175–181.

Marosi, S. 1967. Megjegyzések a magyarországi futóhomok területek genetikájához és morfológiájához. *Földrajzi Közlemények* 15, 231–255.

Mauz, B., Bode, T., Mainz, H., Blanchard, W., Hilger, R., Dikau, R., Zöller, L. 2002. The luminescence dating laboratory at the University of Bonn: equipment and procedures. *Ancient TL* 20, 53–61.

Murray, A. S., Wintle, A. G. 2000. Luminescence dating of quartz using an improved single-aliquot regenerative-dose protocol. *Radiation Measurements* 32, 57–73. DOI: 10.1016/s1350-4487(99)00253-x

Nott, J., Bríant, E., Price, D. 1999. Early-Holocene aridity in tropical northern Australia. *The Holocene* 9 (2), 231–236. DOI: 10.1191/095968399673789264

Nyári, D., Kiss, T. 2005a. Homokmozgások vizsgálata a Duna-Tisza közén. *Földrajzi Közlemények* 129 (3-4), 133–147.

Nyári, D., Kiss, T. 2005b. Holocén futóhomok-mozgások Bács-Kiskun megyében régészeti leletek tükrében. *Cumania*, 83–94.

Nyári, D., Kiss, T., Sipos, Gy., Knipl, I., Wicker, E. 2006a. Az emberi tevékenység tájformáló hatása: futóhomok-mozgások a történelmi időkben Apostag környékén. A táj változásai a Kárpát-medencében. Település a tájban konferencia kiadványa, 170–175.

Nyári, D., Kiss, T., Sipos, Gy. 2006b. Történeti időkben bekövetkezett futóhomok-mozgások datálása lumineszcenciás módszerrel a Duna-Tisza közén. III. Magyar Földrajzi Konferencia CD kiadvány

Nyári, D., Kiss, T., Sipos, Gy. 2007a. Investigation of Holocene blown-sand movement based on archaeological findings and OSL dating, Danube-Tisza Interfluve, Hungary www.journalofmaps.com

Nyári, D., Rosta, Sz., Kiss, T. 2007b. Multidisciplinary analysis of an archaeological site based on archaeological, geomorphological investigations and optically stimulated luminescens (OSL) dating at Kiskunhalas on the Danube-Tisza Interfluve, Hungary. Abstracts book, EAA 142–143.

Olson, C.G., Nettleton, W.D., Porter, D.A., Brasher, B.R. 1997. Middle Holocene Aeolian activity on the High Plains of west-central Kansas. *The Holocene* 7 (3), 255–261. DOI: 10.1177/095968369700700301

Persaits, G., Gulyás, S., Sümegi, P., Imre, M. 2008. Phytolith analysis: environmental reconstruction derived from a Sarmatian kiln used for firing pottery. In Szabó, P., Hédl, R. (eds.) Human Nature: Studies in Historical Ecology and Environmental History. Institute of Botany of the Czech Academy of Sciences, Pruhonice, 87–98.

Rácz, L. 2006. A Kárpát-medence éghajlattörténete a közép- és koraújkorban. In Gyöngyösy, M. (ed.): Magyar középkori gazdaság- és pénztörténet. Jegyzet és forrásgyűjtemény. Bölcsész Konzorcium, Budapest, 34–35.

Rowling, J.E., Fredlund, G.G., Mahan, S. 2003. Aeolian cliff-top deposits and buried soils in the White Reiver Badlands, South Dakota, USA. *The Holocene* 13 (1), 121–129.

Shulmeister, J. 1992. A Holocene pollen record from lowland tropical Australia. *The Holocene* 2, 107–16.
DOI: 10.1177/095968369200200202

Shulmeister, J., Lees, B.G. 1992. Morphology and chronostratigraphy of a coastal dunefield; Groote Eylandt, northern Australia. *Geomorphology* 5, 45–53. DOI: 10.1016/0169-555x(92)90023-h

Sipos, Gy., Kiss, T., Nyári, D. 2006. OSL mérés lehetőségei. Homokmozgások vizsgálata Csengele területén. Environmental Science Symposium Abstracts, Budapest, 43-45.

Stokes, S., Swinehart, J.B 1997. Middle- and late-Holocene dune reactivation in the Nebraska Sand Hills, USA. *The Holocene* 7 (3), 263–272. DOI: 10.1177/095968369700700302

Sümegi, P. 2001. A Kiskunság a középkorban – geológus szemmel In Horváth, F. (ed) A csengelei kunok ura és népe. Archaeolingua Kiadó, Budapest, 313–317.

Sümegi, P., Lóki, J. 1990. A lakiteleki téglagyári feltárás finomrétegtani elemzése. *Acta Geographica Debrecina 1987-1988, Tomus* 26-27, 157–167.

Sümegi, P., Lóki, J., Hertelendi, E., Szöőr, Gy. 1992. A tiszaalpári magaspart rétegsorának szedimentológiai és sztatigráfiai elemzése. *Alföldi Tanulmányok* 14, 75–87.

Sümegi, P. 2005. Loess and Upper Paleolithic environment in Hungary. An Introduction to the Environmental History of Hungary. Aurea Kiadó, Nagykovácsi, 183–211.

Ujházi, K. 2002. A dunavarsányi garmadabucka fejlődéstörténete radiometrikus kormeghatározások alapján. *Földtani Közlemények* 132 (különszám), 175–183.

Ujházi, K., Gábris, Gy., Frechen, M. 2003. Ages of periods of sand movement in Hungary determined: through luminescence measurements. *Quaternary International* 111, 91–100.

Wicker, E. 2005. Excavation documentation of Apostag Szilas kelet Katona József Museum, Kecskemét.

Wilkins, D.E., Currey, D.R. 1999. Radiocarbon chronology and δ13 C analysis of mid- to late-Holocene Aeolian environment, Guadalupe Mountains National Park, Texas, USA. *The Holocene* 9 (3), 363–371. DOI: 10.1191/095968399677728249

Wilson, P., Braley, S.M. 1997. Development and age structure of Holocene coastal sand dunes at Horn Head, near Dunfanaghy, Co Donegal, Ireland. *The Holocene* 7 (2), 187–197. DOI: 10.1177/095968369700700206

Wintle, A.G., Clarke, M.L., Musson, F.M., Orford, J.D., Devoy, R.J.N 1998. Luminescence dating of recent dunes on Inch Spit, Dingle Bay, southwest Ireland. *The Holocene* 8 (3), 331–339.
DOI: 10.1191/095968398671791976

JOURNAL OF ENVIRONMENTAL GEOGRAPHY
Journal of Environmental Geography 7 (1–2), 49 – 56.
DOI: 10.2478/jengeo-2014-0006
ISSN: 2060-467X

EFFECT OF CLIMATE CHANGE ON THE HYDROLOGICAL CHARACTER OF RIVER MAROS, HUNGARY-ROMANIA

György Sipos*, Viktória Blanka, Gábor Mezősi, Tímea Kiss, Boudewijn van Leeuwen

Department of Physical Geography and Geoinformatics, University of Szeged, Egyetem u. 2-6, H-6722 Szeged, Hungary
*Corresponding author, e-mail: siposgy@geo.u-szeged.hu

Research article, received 28 February 2014, accepted 13 March 2014

Abstract

It is highly probable that the precipitation and temperature changes induced by global warming projected for the 21st century will affect the regime of Carpathian Basin rivers, e.g. that of River Maros. As the river is an exceptionally important natural resource both in Hungary and Romania it is necessary to outline future processes and tendencies concerning its high and low water hydrology in order to carry out sustainable cross-border river management. The analyses were based on regional climate models (ALADIN and REMO) using the SRES A1B scenario. The modelled data had a daily temporal resolution and a 25 km spatial resolution, therefore beside catchment scale annual changes it was also possible to assess seasonal and spatial patterns for the modelled intervals (2021-2050 and 2071-2010). Those periods of the year are studied in more detail which have a significant role in the regime of the river. The study emphasizes a decrease in winter snow reserves and an earlier start of the melting period, which suggest decreasing spring flood levels, but also a temporally more extensive flood season. Changes in early summer precipitation are ambiguous, and therefore no or only slight changes in runoff can be expected for this period. Nevertheless, it seems highly probable that during the summer and especially the early autumn period a steadily intensifying water shortage can be expected. The regime of the river is also greatly affected by human structures (dams and reservoirs) which make future, more detailed modelling a challenge.

Keywords: River Maros, catchment hydrology, climate change, RCMs

INTRODUCTION

River Maros has always been an important natural resource on the Southern Great Plains of the Carpathian Basin. The amount of water it drains annually equals to the total water consumption of Hungary. Although only part of this water is utilised, the river is by far the most significant water resource for irrigation and industrial activity in the region. Besides, it also feeds a thriving riparian ecosystem and has a unique geomorphological character. The availability and quality of its resources are endangered by several factors. From among these the short and long term effects of human interventions and that of climate change have to be emphasized. The future of River Maros and adjacent territories is determined basically by the amount of water drained by the river.

Previous research has demonstrated that the Late Pleistocene and Holocene evolution of the river has primarily been affected by climatic variations (Kiss et al., 2013). The fluvial system is still very active and it is highly sensitive to external forcing factors (Kiss and Sipos, 2007), from which recent climate change is getting to be more and more pronounced. In the near future climate change will supposedly alter the duration and pattern of both flood and low water discharges and the intensity of channel development. Large number of investigations have demonstrated that the actual changes of the temperature and precipitation will have significant effects on all factors of the environment, and can also alter the rate of geomorphologic, and especially fluvial processes (Dikau and Schrott, 1999).

The possible causes of climate change concerning river hydrology and fluvial activity has been addressed by several studies recently. These apply for the simulations climate models to predict future deviations in temperature and precipitation. For concluding general trends on large catchments global climate models (GCM) are applied by several studies (e.g. Boyer et al., 2010; Chung and Jung, 2010; Zeng et al., 2012), however, differences in the topography and hydrology of subcatchments would call for the downscaling of global models (Dobler et al., 2012), or the application of regional climate models (RCM) at best (Veijalainen et al., 2010). Studies agree, however that the application of several models and emission scenarios can increase the reliability of results (Kay et al., 2006; Smith et al., 2013). However, moderate analyses usually apply the A1B SRES scenario for calculations.

One of the most important problems addressed by hydrologists is the expectable change in flood frequency and flood height. In this respect variations in snow/precipitation ratio and seasonal rainfall distribution are key parameters (Boyer et al., 2010; Bell et al., 2012). As a consequence of these flood hazard can increase even if total annual runoff is expected to decrease (Kay et al., 2006; Zeng et al., 2012). Furthermore, in several cases the elongation of the flood season, can result a lower predictability concerning flood timing in the future (Dobler et al., 2012).

On the other end water shortage may pose serious problems for water managers (Lorenzo-Lacruz et al., 2010; Koutroulis et al., 2013). Beside well known consequences, such as future limitations in water use and related agricultural, economic and social conflicts, even accelerated geomorphological change can be expected at certain catchments (Smith et al., 2013). Concerning low stage periods predictability of river hydrology is highly limited by human interventions (e.g. water retention, irrigation), which is a problem on most of engineered rivers (Kiss and Blanka, 2012). Among these circumstances adaptation will be a key issue.

In this study a conceptual framework was set up for the trends in catchment scale hydrological changes concerning River Maros. The major questions were: (1) What changes can be expected in temperature and precipitation on the watershed of the river in the 21st century? How these changes can affect the regime of the river and the total annual volume of water drained? The tendencies of future changes were explored by using regional climate models. The study aims to provide base for future hydrological modelling and runoff calculations.

STUDY AREA

The catchment of River Maros is located in the south-eastern part of the Carpathian Basin. 92% of its total area (30 000 km2) belongs to Romania, the remaining 8% to Hungary (*Fig.1*). River Maros and its tributaries are mostly fed by precipitation and overland flow. Due to the geology of the catchment (overwhelmingly volcanic and crystalline rocks) and the high proportion of very steep slopes floods rise relatively quickly, and last for only a short time. Two major floods may develop annually on the river. The first is due to snowmelt in early spring, the second is caused by early summer rainfall usually in June (*Fig. 2*). Following the April-June floods the rest of the year is characterized by low stages, which last for approximately 10 months from June till March, with a minimum water delivery at October (Boga and Nováky 1986).

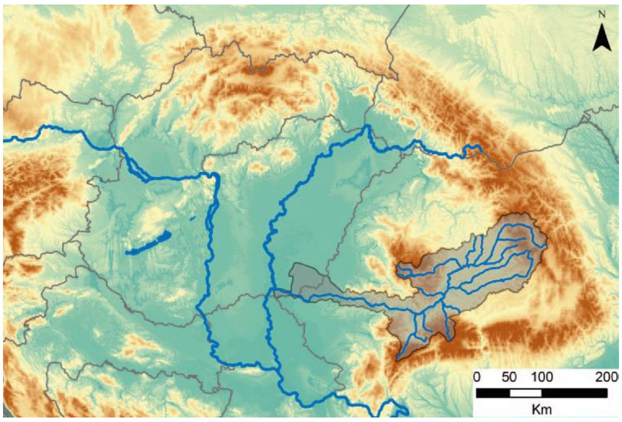

Fig. 1 The location of the Maros catchment within the Carpathian Basin

The mean annual discharge of the river is 160 m^3/s. Peak and minimum discharges are around 2000-2500 m^3/s and 30-50 m^3/s, respectively.

The area is dominated by westerly winds, though from time to time the effect Eastern-European and Mediterranean air masses are also significant (Csoma 1975). The annual mean temperature on the catchment is between 4–11 °C, though there is a great territorial variation, determined primarily by the topography. As a general rule the value of the mean annual temperature is increasing continuously in an east-west direction. In the Giurgeu Mountains the mean annual temperature in the 20th c. has been 4–6 °C, in the Transylvanian Basin 8–9 °C, west of the Arad–Oradea line it has been just above 10 °C, while southwest of the Nadlac–Szeged line it has been above 11 °C (Csoma, 1975; Andó, 2002). 70–80% of the water drained by the river is originating from precipitation (Andó, 1993, 2002). Still, there is no close relation between floods and the temporal distribution of precipitation (Andó, 2002), as the greatest floods are initiated by the melting of snow, accumulating in the winter period. The spatial distribution of precipitation shows a great variation (*Fig. 3*). At the source of the river annual values are around 600 mm, going downstream this amount can double on the western slopes of the Gurghiu Mountains,but reaching the closed Transylvanina basin it falls back to 600 mm again. It increases again in the region of the Sebeş and Retezat Mountains. West of Lipova precipitation decreases continuously (Csoma, 1975).

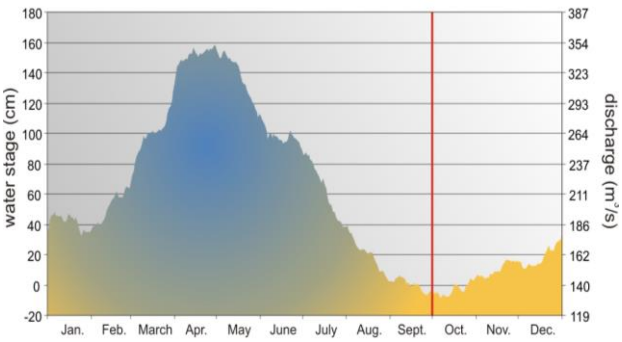

Fig. 2 An average hydrological year based on 50 year mean values (1950-2000)

Climate model simulations for the 21st century predict a continuous, but uneven temperature rise, with the most intense increase occurring in the summer months in the Carpathian Basin. The change in total annual precipitation in the models is not significant; however, the temporal distribution is expected to become more uneven: decreasing summer and increasing winter precipitation (Bartholy et al., 2008, Szabó et al., 2011; Csorba et al., 2012). Climate simulations do also emphasize that extreme weather events may occur more frequently in the next century. This will be especially true for drought periods which will be longer and more severe than before (Szépszó et al., 2008).

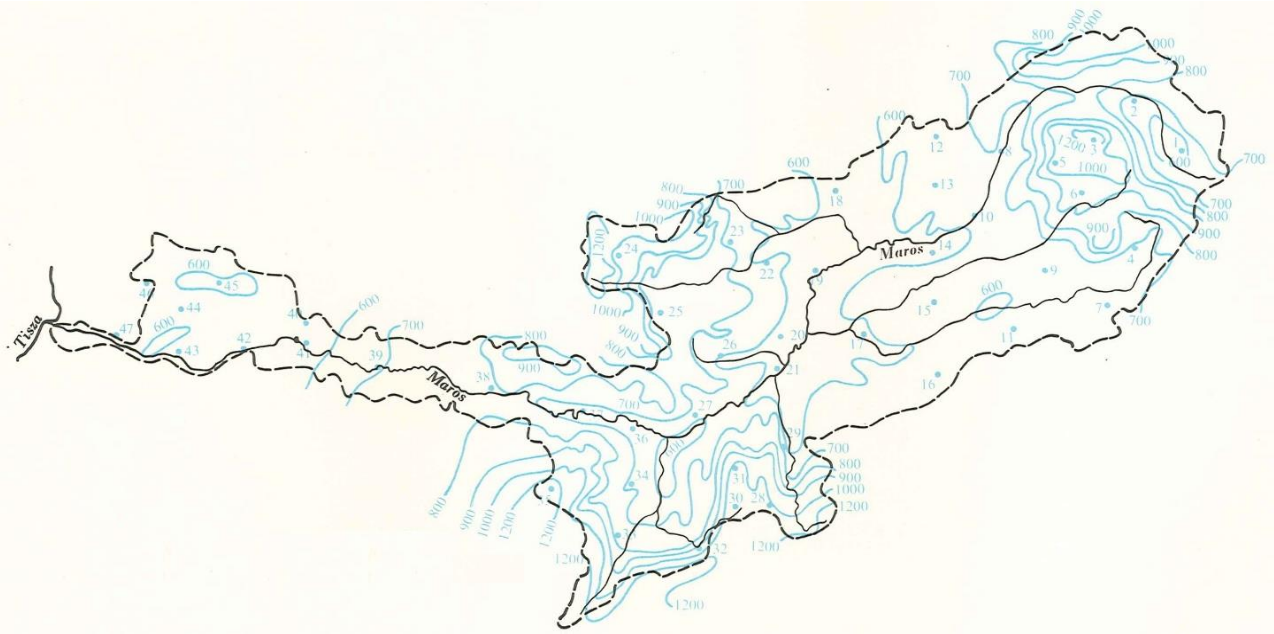

Fig. 3 Yearly mean precipitation (1901-1940) on the Maros catchment (Csoma, 1975)

METHODS

Climatic changes concerning the entire Earth are best predicted by global numerical models. These incorporate the most important processes and relationships acting in and between the major elements of the Earth system (atmosphere, oceans, continents, ice sheets, biosphere, society). The horizontal resolution of global models, however, is only around 100 km, which does not provide adequate information for performing smaller scale, regional analyses (Szépszó and Zsebeházi, 2011).

For the investigation of smaller areas, such as the drainage basin of a river, regional climate models can be applied. The resolution of these is much better than that of global models, since input data are more detailed and smaller scale relationships can be considered. As a consequence, atmospheric processes and surface changes can be predicted more precisely for a given area (van der Linden and Mitchell, 2009; Szabo et al., 2011). Large, Earth scale models, however determine the possible end values of the regional forecast (Giorgi and Bates, 1989; Giorgi 1990). The projection of climate processes to the future includes several uncertainties. These are caused by the natural oscillation of the climatic system, the complicated relationships between environmental elements, the limitations concerning the resolution of input data and the hardly predictable social-economic processes (Cubasch et al., 2001; Hawkins and Sutton, 2009). From among the factors above mankind can affect mostly social-economic processes, and most of all, these factors can significantly determine the rate of climate change.

Several regional climate models comprise the territory of the Carpathian Basin, these are the ALADIN, the REMO, the PRECS and the RegCM (Szépszó et al., 2008). For the purposes of estimating future climatic processes on the Maros catchment, the change of climatic parameters was calculated on the basis of the ALADIN (www.cnrm.meteo.fr/aladin) and the REMO (www.remo-rcm.de) models, since these are based on the SRES A1B scenario.

These models assume a moderate increase in the emission of greenhouse gases and an average degree of global warming. Concerning population numbers the A1B scenario assumes an increase till the middle of the century and a decrease later, besides, it foresees a fast economic growth, the quick spread of new and more efficient technologies and a balance between the use of fossil and renewable energy sources (Nakicenovic and Swart, 2000; IPCC 2007).

The horizontal resolution of the applied model data is 0.22° (approximately 25 km). The climate projections were generated by the Numerical Modelling and Climate Dynamics Division of the Hungarian Meteorological Service. For making comparisons, temperature and precipitation data were investigated. Expected changes were examined for those periods of the year which are the most important in terms of floods and low water periods.

The models provide daily temperature and precipitation data for the 2021–2050 and the 2071–2100 periods. Results are given as the difference from the daily average values of the 1961–1990 reference period in mm for precipitation and in °C for temperature. From the daily data series average values were calculated for the two future periods (2021-2050 and 2071-2100) for those months which are important in terms of the hydrology of River Maros. Values were calculated for the grid points, interpolation between points was made by using kriging.

At present our aim was to demonstrate the tendency of changes, but later on the basis of these data further models can be generated in terms of runoff and water balance.

RESULTS

Compared to the reference values both models predict an average 1.3–1.4 °C temperature increase for the 2021–2050 period in the winter months (*Fig. 4*). However, in a longer perspective the models forecast somehow different values. According to the REMO the rise of mean temperature can be as much as 3.9 °C for 2071–2100, which is a substantial increase. The ALADIN predicts a little lower increase, being around 2.1 °C *(Fig. 4*, *Table 1).* However, even if we take the more optimistic version a significant warming can be expected on the entire catchment. If we take a look at *Fig. 4*, in the first modelling period temperature rise seems to be uniform on the catchment. However, later the Eastern Carpathian tributaries and the lowland section of the river can be more affected. Concerning the entire catchment, average precipitation values calculated by the two models are very different. According to the ALADIN, practically no change can be expected, while the REMO predicts a 22 mm increase for 2021–2050 and 34 mm for 2071–2100 (*Fig. 5*, *Table 2*). Average values though hide some regional differences. Both models agree that there can be a notable increase in precipitation on the eastern mountainous part of the catchment (Giurgeu Mountains), while only a slight increase (REMO) or even a substantial decrease (ALADIN) can be expected in the west (*Fig. 5*). Based on the above, it seems well supported that due to general warming the average snow reserve will decrease on most of the sub-catchments. Although, at higher altitudes in the Eastern Carpathians the snow/precipitation ratio might be higher as a matter of precipitation increase. The severity of flooding on these sub-catchments will mostly be determined by the intensity of snowmelt.

Table 1 Average temperature change (°C) on the Maros catchment

	REMO		ALADIN	
	2021-2050	2071-2100	2021-2050	2071-2100
December-February	1.4	3.9	1.3	2.1
May-June	1.1	2.4	1.5	3.1
July-August	1.4	5.0	3.0	5.5
September-October	2.2	4.8	2.5	4.6
Annual	1.4	3.8	2.02	3.55

March and April temperature has a significant effect on the start and intensity of snow melt and therefore the development of floods. Based on the models, a general temperature increase can be expected (*Fig.*

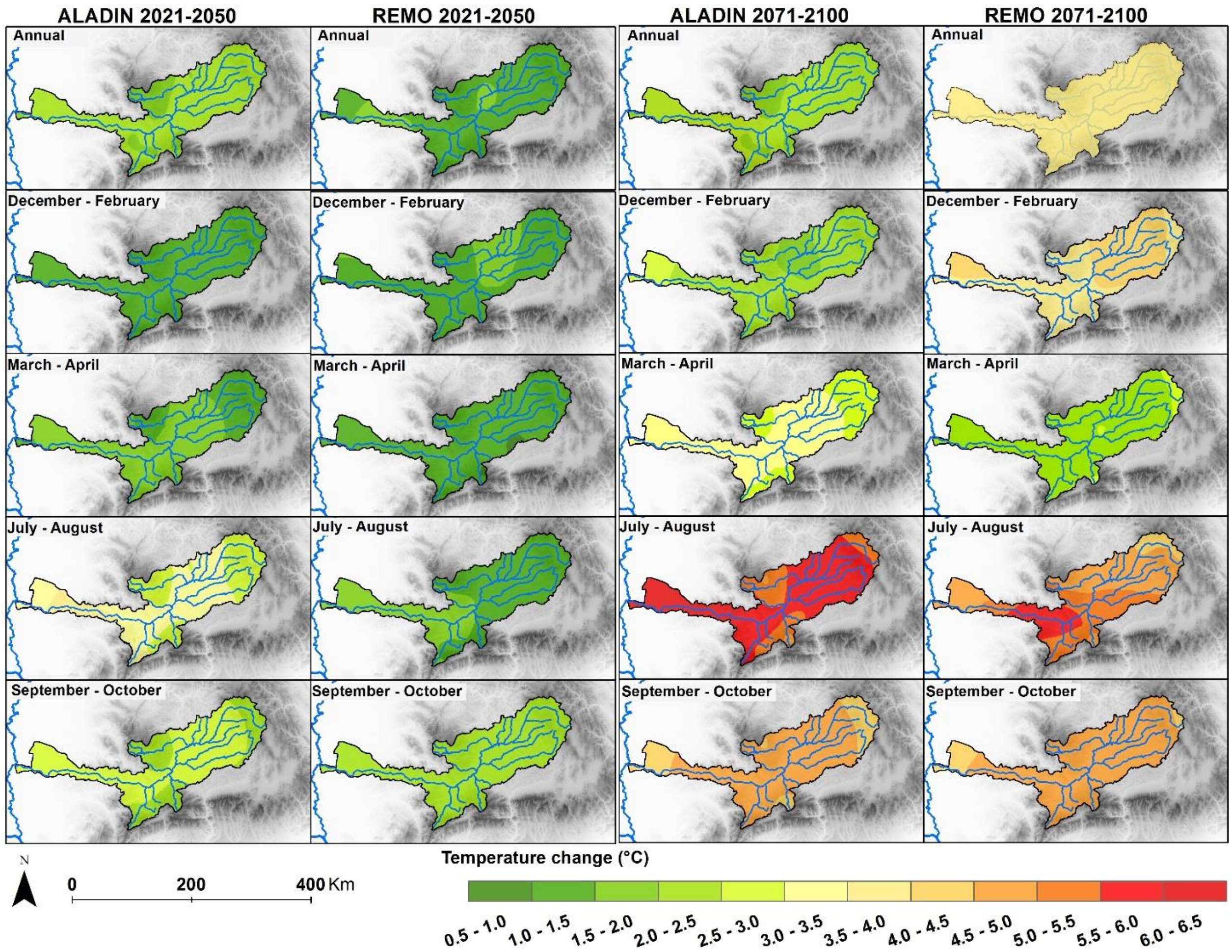

Fig. 4 Model predictions of temperature change compared to the reference period

4, *Table 1*). For the first period (2021–2050) an average 1.1–1.5 °C growth is suggested. By the second period (2071–2010) this value can be as much as 2.4 °C (REMO) or 3.1 °C (ALADIN). Warming up can be more intensive in the Transylvanian Basin and on the lowlands, however the Eastern catchment can also face an approximately 1.0 °C later a 2.0 °C temperature increase in the spring period during the 21st century (*Fig. 4*, *Table 1*). These changes suggest that in an average year early spring snowmelt can be faster in the upland catchment. This does not necessarily mean greater floods, because we have seen the total snow reserve can be slightly lower. Nevertheless, the period of flood development will extend, and in years of higher winter precipitation the chance of the development of extreme floods can increase.

Based on previous observations, the second potential flood of the year can occur as a consequence of May and June rainfalls (Andó, 2002). Model predictions are ambiguous in this respect (*Fig. 5*, *Table 2*).

According to average data calculated for the entire catchment, REMO forecasts an insignificant change in early summer rainfall for the 2021–2050 period, and a substantial 50 mm decrease for 2071–2100. On the other hand ALADIN predicts an approximately 30 mm increase for the first period and just a minor increase for the second (*Fig. 5*, *Table 2*). The pattern of change is also different. According to the ALADIN model, the most significant increase can be expected in the middle of the catchment. On the contrary REMO heralds the most significant decrease also to this area (*Fig. 5*). Therefore, the direction of change in this case is highly uncertain. Consequently, it is hard to tell whether the significance of early summer rain-fed floods will increase or decrease. If we take the average of the two models rather insignificant changes can be expected.

Table 2 Average precipitation change (mm) on the Maros catchment

	REMO		ALADIN	
	2021-2050	2071-2100	2021-2050	2071-2100
December-February	22.38	34	0.12	-5.5
May-June	-3.5	-50.1	32.6	35.4
July-August	-0.3	-55.3	-21.5	-55.6
September-October	2.2	8.3	-2.8	-17.4
Annual	2	-53	44	-20

As it was mentioned earlier, the low water period starts in July (Boga and Nováky, 1986). From August discharges can be as low as 50 m^3/s. The

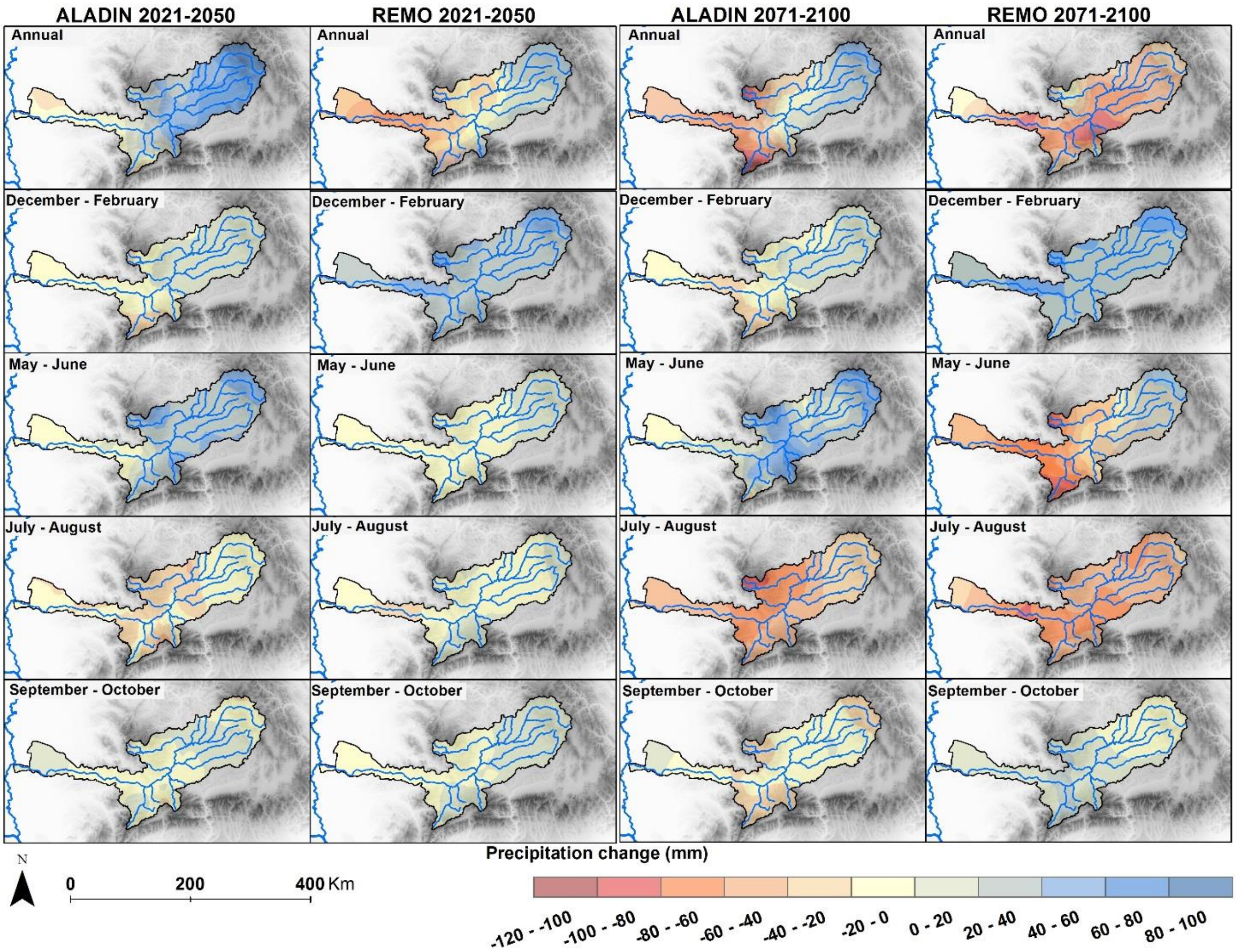

Fig. 5 Model predictions of precipitation change compared to the reference period

volume of water arriving to the lowland sections is greatly determined by the intensity of evaporation on the catchment. Of course, human interventions, such as water storage, can also be of great significance. In this respect July–August temperatures are very important. Both models forecast an increase, being between 1.4 °C (REMO) and 3.0 °C (ALADIN), for the first modelling period (2021–2050). Concerning the second period (2071–2100) the increase can be even more significant, reaching 5.0 °C (REMO) or 5.5 °C (ALADIN) (*Fig. 4, Table 1).* The expected temperature growth is fairly uniform on the catchment, however, according to ALADIN, warming will mostly affect the Gurghiu Mountains and the lowland areas, while REMO predicts the most intensive increase on the middle part of the catchment. Therefore, the spatial pattern for warming cannot be unambiguously determined. In the meantime, there is a high chance for the decrease of summer precipitation. Regarding the average values for the entire catchment the ALADIN model forecasts a 20 mm decrease for 2021–2050, while according to the REMO, catchment averages may not change. However, both models agree that by 2071–2100 precipitation loss can be around 55 mm (*Table 2*), affecting mostly the middle part of the drainage basin. The decrease can be less intensive in the western slopes of the Hargitha, Giurgeu and Gurghiu Mountains (*Fig. 5*). Concerning the summer period, therefore, increasing evaporation and decreasing precipitation can be forecasted. This can lead to a significant reduction in average discharges, which may result an increasing water shortage during the low water period starting from July–August.

Based on previous observations (Konecsny and Bálint, 2007), usually the September–October period brings the lowest discharges on River Maros (sometimes only 30–40 m^3/s). If catchment scale average temperature change is considered the two models are in good agreement. For 2021–2050 both models forecast a temperature increase, being around 2–3 °C, while between 2071–2100 average warming can be as much as 4–5 °C (*Fig. 4*, *Table 1*). Concerning the spatial distribution of temperatures, warming might affect less the slopes of the Apuseni Mountains and the Gurghiu Mountains, but in the Transylvanian Basin and the Tarnava Tableland temperature rise can be dramatic (*Fig. 4*). Precipitation change is less obvious on the basis of the models, and catchment scale averages seem to stay more or less the same as the 1961–1990 reference values (*Table 2*). The calculated few mm changes are insignificant and they are within the error of the prediction.

The high correspondence of the two models suggests that there is going to be a significant warming in the early autumn period. In the meantime average precipitation values will hardly change, which can finally result a more intensive water loss through evaporation. This can lead to the development of long drought periods along River Maros.

DISCUSSION

Although we have seen that in certain cases the two models do not reinforce each other, there are some clearly recognisable tendencies in terms of future climate. Warming will be general both in spatial and temporal terms. However, lower lying closed areas, such as the Transylvanian Basin, can be more severely affected. It seems also clear that temperature rise will be the most significant in the summer–autumn period, though the REMO model forecasts significantly warmer winters as well. In the meantime, changes in precipitation are harder to predict. What seems obvious though is that the late summer period will face a significant precipitation decrease on the basis of average values. Changes in other seasons are less unambiguous (*Table 3*).

Table 3 Precipitation change (%) compared to the reference period on the mountain section of the catchment

	REMO		ALADIN	
	2021-2050	2071-2100	2021-2050	2071-2100
December-February	21%	31%	0%	-8%
May-June	-2%	-26%	16%	17%
July-August	0%	-32%	-12%	-32%
September-October	2%	8%	-3%	-17%
Annual	0%	-7%	6%	-3%

If annual mean values are considered, a significant 1.4 °C (REMO) and 2.0 °C (ALADIN) temperature increase can be predicted already for 2021–2050. Moreover, by 2071–2100 overall warming can be 3.6 °C (ALADIN) and 3.8 °C (REMO) compared to the values of the 1961–1990 reference period. Interestingly, annual precipitation values show a slight increase for 2021–2050 in case of the ALADIN model, but for the 2071–2100 period both models forecast a significant, 20–50 mm decrease. Taking into consideration that the average precipitation is between 600–1000 mm on the catchment, this means a 5–10% reduction in annual runoff. The decrease can be even more significant if increasing evaporation is accounted, but further modelling is necessary to explore these relationships.

Concerning the hydrological regime of the river we can expect a more uniform runoff during the winter, however, early spring snow melt can be more intensive. In the meantime early summer floods might be less significant (*Table 3*). Therefore, the frequency and average magnitude of floods will slightly decrease, however, if conditions are suitable (high winter precipitation and fast snowmelt) extreme floods can of course occur. Although several climate-related studies emphasize the relevance of high-precipitation extremes (Szépszó et al., 2008), these will be characteristic mostly on the western half of the Carpathian Basin (Horányi et al., 2009). Consequently, from a climatic aspect the hazards and conflicts related to floods and flood protection will not increase significantly along the Maros/Mureş in the near future. On the other hand, results show that summer and autumn low water extremes may be more frequent, and severe water shortage may occur along the lower section of the river from time to time (*Table 3*). Moreover, as we have seen, total annual runoff will certainly decrease in the long run. According to Konecsny (2010), there are already periods with significant water deficit, meaning that the discharge is lower than the statistically determined average low water value. Thus, the main problems and conflicts related to the changing regime of the river will be related primarily to low water events.

CONCLUSIONS

Calculations concerning the future climate of the Maros/Mureş catchment were outlined, and the tendency of expectable changes was assessed in this study concerning the hydrological regime of the river.

Due to increasing temperatures at winter the average snow reserve can decrease on several subcatchments. Nevertheless, at higher altitudes greater reserves may develop, since models herald a slight increase in winter precipitation. Spring snowmelt can be faster in the upland catchment, thus in years when winter precipitation is high the probability of extreme floods can increase. In general, however, the magnitude of floods is expected to decrease. Based on the models, considerable changes in the volume of early summer rainfed floods are not expected. For the summer and early autumn period dramatically increasing temperature and decreasing precipitation can be forecasted. This can lead to a significant reduction in average discharges. On a catchment scale mean annual temperature is expected to increase by 1.4-2.0 °C and 3.6-3.8 °C in average by 2021–2050 and 2071–2100, respectively. Mean annual precipitation presumably will only slightly change by the first modelling period, however, for 2071–2100 the models forecast a significant, 20–50 mm decrease. Considering the above a 5–10% reduction can be expected in annual runoff, and the severity of droughts will certainly increase.

Consequently, the main problems and conflicts of the future will be related primarily to low water events. Industrial, agricultural, ecological and recreational demands need to be harmonised as each of these will grow during the increasingly hot and dry summer period. All these problems call for a unified water management strategy with a sustainable share of resources between the upstream and lowland sections of the river and also between the two neighbouring countries.

Acknowledgements

The research was supported by the HU-RO/0901/266/2.2.2, HUSRB/1203/121/130 cross-border projects, and the Hungarian Research Fund (OTKA 100761).

References

Andó, M. 1993. The geography of the Mures River. *Acta Geographica Szegediensis* 31, 1–9.

Andó, M. 2002. A Tisza vízrendszer hidrogeográfiája. SZTE Természeti Földrajzi Tanszék, Szeged.

Bartholy, J., Pongrácz, R., Gelybó, Gy., Szabó P. 2008. Analysis of expected climate change in the Carpathian Basin using the PRUDENCE results. *Időjárás Quarterly Journal of the Hungarian Meteorological Service* 112, 249–264.

Bell, V.A., Kay, A.L., Cole, S.J., Jones, R.G., Moore, R.J., Reynard, N.S. 2012. How might climate change affect river flows across the Thames Basin? An area-wide analysis using the UKCP09 Regional Climate Model ensemble. *Journal of Hydrology* 442–443, 89–104. DOI: 10.1016/j.jhydrol.2012.04.001

Boga, L., Nováky, B. 1986. Magyarország vizeinek műszaki-hidrológiai jellemzése: Maros. Vízgazdálkodási Intézet, Budapest.

Boyer, C., Chaumont, D., Chartier, I., Roy, A.G. 2010. Impact of climate change on the hydrology of St. Lawrence tributaries. *Journal of Hydrology* 384, 65–83. DOI: 10.1016/j.jhydrol.2010.01.011

Chang, H., Jung, I.W. 2010. Spatial and temporal changes in runoff caused by climate change in a complex large river basin in Oregon. *Journal of Hydrology* 388 (3–4), 186–207. DOI : 10.1016/j.jhydrol.2010.04.040

Cubasch, U., Meehl, G., Boer, G., Stouffer, R. , Dix, M., Noda, A., Senior, C., Raper, S., Yap, K. 2001. Projections of Future Climate Change. In. Houghton, J.T., Ding, Y., Griggs, D.J., Noguer, M., Van der Linden, P.J., Dai, X., Maskell, K., Johnson, C.A. (eds.). Climate Change 2001: The Scientific Basis: Contribution of Working Group I to the Third Assessment Report of the Intergovernmental Panel. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA. 525–582.

Csoma, J. 1975. A Maros hidrográfiája. In. Vízrajzi Atlasz Sorozat 19 Maros.VITUKI, Budapest. 7–12.

Csorba, P., Blanka, V., Vass, R., Nagy, R.,Mezősi, G. 2012. Hazai tájak működésének veszélyeztetettsége új klímaváltozási előrejelzés alapján. *Földrajzi Közlemények* 136(3), 237–253.

Dikau, R., Schrott, L. 1999. The temporal stability and activity of landslides in Europe with respect to climatic change (TESLEC): main objectives and results.*Geomorphology* 30, 1–12. DOI: 10.1016/S0169-555X(99)00040-9

Dobler, C., Bürger, G., Stötter, J. 2012. Assessment of climate change impacts on flood hazard potential in the Alpine Lech watershed. *Journal of Hydrology* 460–461, 29–39. DOI : 10.1016/j.jhydrol.2012.06.027

Giorgi, F. 1990. Simulation of regional climate using a limited area model nested in a general circulation model. *Journal of Climatology* 3, 941–963. DOI : 10.1175/1520-0442(1990)003<0941:SORCUA>2.0.CO;2

Giorgi, F., Bates, G., 1989. The Climatological Skill of a Regional Model over Complex Terrain. *Monthly Weather Review* 117, 2325–2347. DOI :10.1175/1520-0493(1989)117<2325:TCSOAR>2.0.CO;2

Hawkins, E., Sutton, R., 2009. The potential to narrow uncertainty in regional climate predictions. *Bulletin of American Meteorological Society* 90, 1095–1107. DOI : 10.1175/2009BAMS2607.1

Horányi, A., Csima, G., Szabó, P., Szépszó, G. 2009. Regionális klímamodellezés az Országos Meteorológiai Szolgálatnál. MTA előadás 2009.09.15. (http://www.met.hu/doc/tevekenyseg/klimamodellezes/MTA-2009.09.15.pdf)

IPCC 2007. Climate Change 2007: The Physical Science Basis. Working Group I Contribution to the Fourth Assessment Report of the IPCC. Edited by S. Solomon, D. Qin, M. Manning,Z. Chen, M. Marquis, K.B. Averyt, M. Tignor, H.L. Miller. Intergovernmental Panel on Climate Change, Cambridge University Press, New York, p. 996 (http://www.ipcc.ch)

Kay, A.L., Jones, R.G., Reynard, N.S. 2006. RCM rainfall for UK flood frequency estimation. II. Climate change results. *Journal of Hydrology* 318, 163–172. DOI: 10.1016/j.jhydrol.2005.06.013

Kiss, T., Blanka V. 2012. River channel response to climate- and human-induced hydrological changes: case study on the meandering Hernád River, Hungary. *Geomorphology* 175–176, 115–125. DOI: 10.1016/j.geomorph.2012.07.003

Kiss, T, Sipos, Gy. 2007. Braid-scale geometry changes in a sand-bedded river: Significance of low stages. *Geomorphology* 84, 209–221. DOI: 10.1016/j.geomorph.2006.01.041

Kiss, T, Sümeghy, B., Sipos, Gy. 2013. Late Quaternary paleodrainage reconstruction of the Maros River alluvial fan. *Geomorphology* 204, 49–60. DOI: 10.1016/j.geomorph.2013.07.028

Konecsny, K. 2010: A kisvizek főbb statisztikai jellemzői a Maros folyó alsó szakaszán. *Hidrológiai Közlöny* 90(1), 45–55.

Konecsny, K., Bálint, G. 2010. Low water related hydrological hazards along the lower Mureş/Maros river. In Riscuri şi catastrofe, Universitatea „Babeş-Bolyai”. Facultatea de Geografie. Laboratorul de riscuri şi hazarde. Casa Cărţii de Ştiinţă. Cluj-Napoca 8/6 van der Linden P., Mitchell J.F.B. (eds.) 2009. ENSEMBLES: Climate Change and its Impacts: Summary of research and results from the ENSEMBLES project. Met Office Hadley Centre, Exeter, UK. (http://ensembles-eu.metoffice.com/docs/Ensembles_final_report_Nov09.pdf)

Koutroulis, A.G., Tsanis, I.K., Daliakopoulos, I.N., Jacob, D. 2013. Impact of climate change on water resources status: A case study for Crete Island, Greece. *Journal of Hydrology* 479, 146–158. DOI: 10.1016/j.jhydrol.2012.11.055

Lorenzo-Lacruz, J., López-Moreno, J.I., Beguería, S., García-Ruiz, J.M., Cuadrat, J.M. 2010. The impact of droughts and water management on various hydrological systems in the headwaters of the Tagus River (central Spain). *Journal of Hydrology* 386 (1–4), 13–26.

Nakicenovic, N., Swart, R. 2000: Emissions Scenarios. A Special Report of IPCC Working Group III. Cambridge University Press, Cambridge, UK. 570p.

Smith, V.B., David, C.H., Cardenas, M.B., Yang Z.L. 2013. Climate, river network, and vegetation cover relationships across a climate gradient and their potential for predicting effects of decadal-scale climate change. *Journal of Hydrology* 488, 101–109. DOI: 10.1016/j.jhydrol.2013.02.050

Szabó, P., Horányi, A., Kruzselyi, I., Szepszó, G. 2011. Az Országos Meteorológiai Szolgálat regionális klímamodellezési tevékenysége: ALADIN-Climate és REMO. 36. Meteorológiai Tudományos Napok beszámolókötete. Budapest, 87–101.

Szepszó, G., Zsebeházi, G. 2011. Az ENSEMBLES projekt regionális modelleredményeinek alkalmazhatósága Magyarország éghajlatának jellemzésére. 36. Meteorológiai Tudományos Napok beszámolókötete. Budapest, 59–75.

Szepszó, G., Bartholy, J., Csima, G., Horányi, A., Hunyady, A., Pieczka, I., Pongrácz, R., Torma, Cs. 2008. Validation of different regional climate models over the Carpathian Basin. EMS8/ECAC7 Abstracts 5, EMS2008–A–00645.

Veijalainen, N., Lotsari, E., Alho, P., Vehviläinen, B., Käyhkö, J. 2010. National scale assessment of climate change impacts on flooding in Finland. *Journal of Hydrology* 391 (3–4), 323–350. DOI: 10.1016/j.jhydrol.2010.07.035

Zeng, X., Kundzewicz, Z. W., Zhou, J., Su, B. 2012. Discharge projection in the Yangtze River basin under different emission scenarios based on the artificial neural networks. *Journal of Hydrology* 282, 113–121. DOI: 10.1016/j.quaint.2011.06.009