6. Monitoring and modelling possibilities for increasing sustainability

Drought belongs to the most significant environmental hazards, but it is the less known due to its complexity. It can occur on a small area, but it can cause problems also on continental scale. From temporal aspects the phenomena can last for a week, but also for a decade which was confirmed by instrumental records and paleo climatic reconstructions. Due to its complex character the formation of an early warning system is more difficult compared to other hydro-meteorological hazards (Pulwarty and Sivakumar 2014).

Since drought has significant economic, ecological and social consequences, there were some attempts for its monitoring and prediction in the early 1990s (1994 United Nations Convention to Combat Desertification). Later several assessments were born on regional and international level. A new intention is to have an early warning system (Global Drought Information System, Pozzi et al. 2013) where monthly precipitation deficit maps will be available on global level. For Europe, the European Drought Observatory System (EDO) provides information on drought. For SE Europe, the Drought Management Centre for Southeastern Europe (DMCSEE) aimed the development of a regional early warning system.

An early warning system provides information on the extent and rate of the expected effects of a phenomenon (and the possibilities of mitigation, prevention and adaptation). The base of the system is data integration and processing for setting up environmental models. Two types of model exist: numerical and statistical ones. The former calculates data based on modelling (e.g. climate prognosis for 10 days, for maximum 2 weeks). The latter computes expected values based on the occurrence and frequency of data from previous time periods.

Scale is an important question of drought early warning systems. Global/regional systems are not informative for individual users due to their general content and resolution. Plot scale system cannot be achieved due to the information and data on plot-scale are not available; furthermore, uncertainty increases by the decreasing scale. For better resolution (e.g. 1 km^2) detailed knowledge and monitoring of local factors are necessary, that requires field measurements and high resolution remotely sensed data. Thus, our project aimed at the development of monitoring possibilities of environmental factors playing role in the formation of drought using field measurements, processing of remote sensed data and modelling.

During the project several monitoring and modelling methods were assessed to set up a possible drought monitoring/early warning system:

- Climate parameters: data collection by the procured field meteorological stations and the EUMETSAT system; calculation and use of drought indices; SPI, PaDI indices
- Soil moisture: data collection by the procured field soil moisture measurement station and the EUMETSAT system
- Hydrology: monitoring of surface waters by multispectral satellite images, automatic discharge measurement facilities and hydrological models (MIKE, HEC-HMS and Budyko), groundwater level monitoring using field sensor system
- Vegetation: EVI and NDVI indices (MODIS, Landsat)

6.1. Potentials of soil moisture

6.1.1 Possibilities of field measurement of soil moisture

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Introduction

Based on the several definitions and classifications of drought, three environmental factors or factor groups are of high importance (Pálfai 2004). Meteorological drought is characterised by classical meteorological parameters (precipitation, air temperature, humidity, wind speed); hydrologic drought by the water level of water flows and lakes and by the groundwater table; and soil drought by soil moisture data measured in different depths. The expected aridity of a given period can be predicted on the basis of meteorological forecasts, therefore, the uncertainty of the prediction increases with the increasing duration of the forecasted period. From agricultural point of view soil moisture is equally important. During severe dry periods in summer the upper 10-20 cm of soil can extremely dry out. If the soil is dry in 70-80 cm depth at the beginning of vegetation period, meaning that soil moisture is below field capacity, the probability of drought occurrence increases. The reason for this is usually the severe lack of precipitation in the former autumn-winter period. This, in itself, does not necessarily lead to summer drought, but if the spring and early summer periods are not wetter than average, serious agricultural damages can be expected, which may also affect perennial plants with deeper root system.

In the WAHASTRAT project 16 hydro-meteorological stations were installed (Fig. 6.1 on page 246) for measuring meteorological parameters, monitoring spatial differences, and the determination of soil drought using soil moisture sensors. This chapter demonstrates the changes of soil moisture in the first half of 2014 in case of some Hungarian stations.

Methods

In Serbia 8 stations have been installed on chernozem, meadow chernozem, solonczak and meadow soils; in Hungary 8 stations on sand, chernozem, alluvial and meadow soils. The stations (Fig. 6.2 on page 247) are suitable for measuring the following parameters:

1. Precipitation in mm – using reed switch
2. Air temperature 2 m above the surface (°C)
3. Air moisture 2 m above the surface (%)
4. Average wind power 2 m above surface (m/s) – using anemometers
5. Wind direction (0° - West)
6. Soil moisture in six different depths (10-20-30-45-60-75 cm). The operation of the applied soil moisture sensor (EC-5) is based on the measuring of the soil dielectric constant, and it is calibrated using a series of samples with controlled moisture values in laboratory. The actual soil moisture is calculated in the percentage of the total volume.

The above listed parameters are measured every hour, and in case of precipitation and wind characteristics the summarised and average values are represented.

During the allocation of the measurement stations it was aimed to cover most of the characteristic soil types of the region; mostly soils of drought-affected cultivated areas were focused, thus e.g. salt-affected soils were excluded from the monitoring. Detailed soil investigations were carried out in case of all selected stations; all soil layers and the layers of the sensors were sampled and the following laboratory measurements were carried out:

a) basic pedological investigations on the characteristic levels: plasticity index according to Arany (according to MSZ 08-0205: 1978), pH_{H2O}, salt content, carbonate content (MSZ 08-0206/2: 1978) and humus content (MSZ 21470-52: 1983).
b) soil particle size distribution from the characteristic levels, according to MSZ 08-0205: 1978.
c) hydro-physical characteristics for all sampled depths: bulk density, maximum and field water capacity, actual soil moisture, hygroscopy, wilting point and hydraulic conductivity were determined from the undisturbed soil samples (Stefanovits 1992).

On the basis of measured data soil moisture conditions were described using field capacity and wilting point as a basis of comparison among the hydro-physical characteristics. Those soil levels were regarded favourable in the point of wetness, where soil moisture was max. 5 v/v % less than field capacity. Soil is considered dry if soil moisture is lower than that value; and if moisture is lower than wilting point, soil is extremely dry. Quantification of water scarcity was done by calculating the necessary infiltration in mm to reach field capacity. The upper three sensor values refer to 10-10 cm, and the lower three sensor values to 15-15 cm levels. It means that 1 and 1.5 mm of precipitation should infiltrate for 1 v/v % moisture increase. The water shortage of a soil profile in mm is calculated from the summarized scarcity of the levels. Following this approach, the amount of infiltrated water into a particular depth can be defined in mm after a rainfall event. This method can be used for the determination of "utilized" water in mm from the precipitation depending on soil type and surface temperature.

Results

Öreghegy and Kelebia soil moisture stations (HU01 and HU02) were installed on Arenosol (humic sandy soil), characteristic for the Danube-Tisza Interfluve blown sand area; Kiskundorozsma, Röszke, and Szentes-Fertőd (HU05, HU06 and HU08) stations in the Dél-Tisza valley are located on Chernozems of Dél-Tiszántúl determined by the highest soil fertility; Tápai-rét and Gerencshát (HU03 and HU07) stations are on Fluvisols (humic alluvial soils); and Batida station is on Vertisol (steppic meadow solonetz). Apart from the two Arenosols, all soils show anthropogenic influence, although their original layers still exist: some bricks and concrete parts were found in their topsoil, while strong compaction can be observed even in lower layers in Dorozsma and Batida. The investigation results of the characteristic levels of soils are summarised in Table 6.1 (page 250), and photos of the profiles of the main types are also attached (Fig. 6.2 on page 247).

The hydro-physical features of the soils reflect their genetic types and the experienced compaction (Table 6.2 on page 251). Two disadvantaged features can be highlighted from laboratory data regarding drought sensitivity:

a) It is a well-known fact that high permeability and weak water-storing capacity (low, even 10 v/v % field capacity) of sand profiles lead to the death of cultivated plants on them due to even relatively short hot and dry periods.

b) Due to soil compaction extremely low permeability was experienced in almost all other profiles (except for Röszke) in several depths, usually with high bulk density and wilting point, low maximum water capacity, and gravitational pore space. The topsoil in Batida and Gencshát is especially compacted, where the temporal change of soil moisture in deeper layers proves the existence of compacted levels above them.

During the installation of the stations at the turn of 2013-2014, a considerable soil drought was experienced: the previous dry summer and autumn, furthermore the 0 mm precipitation in December contributed to the soil moisture values 15-20% below field capacity. The most favourable moisture conditions were characteristic of Chernozems, while in the case of sandy soils and clayey hydromorphic soils the situation was much more severe (Table 6.3 on page 253).

The following one-one and a half months brought 40-70 mm of precipitation, owing to which the extremely dry winter condition seemed to cease in most soils; but the clayey, compacted soils in Batida and Gencshát still seriously lacked water at the end of February, too. With the help of soil moisture data soil water shortage of individual profiles were quantified in the investigated period. Due to the physical characteristics of soils, sandy soils filled up rapidly – later they drained and dried out at the same speed –; Chernozems were characterised by slow and balanced changes in water balance, while in case of the least favourable Fluvisol in Gencshát infiltration occurred only in the upper 20-25 cm, and in February 40 mm of precipitation was still missing to reach field capacity (Table 6.3 on page 253).

The following spring months were characterised by various precipitaton. March was drier, April, May and June were wetter, and apart from April, all months had their monthly precipitation during a couple of days, usually in forms of intensive showers. Such periods were, for example, 24th March, 2-3rd May, 13-16th May and 23-25th June. Especially June showed duality, since in the first three weeks there was no significant precipitation. For more detailed analyses Kelebia and Tápai-rét stations were chosen between 14-15th May (Fig. 6.3 on page 250), and Batida and Gencshát stations in June (Fig. 6.4 on page 254).

The fronts of mid-May represent the water management differences between Arenosols and finer textured (and compacted) Fluvisols. Before the rainfall events soil moisture had already been under field capacity in the upper 20 cm in Kelebia, but underneath the value was around or above field capacity. After the latest (3rd May) precipitation of more than 40 mm (10 days previously), considerable drying was experienced only in the upper 30 cm; soil moisture did not change significantly under that during the examined period. The rainfall on 14th May had its effect immediately on the soil: in a couple of hours soil moisture rose in every depth with a temporal shift. The weak water-storing capacity of sand is demonstrated in the breaks of rainfalls or rather 1-2 hours after its end, when soil moisture started to decrease in the upper 30 cm, and water infiltrates to deeper layers.

In the dry period prior to 13th May on the Tápai-rét a decrease can also be seen on the data of the upper three sensors, however, moisture data here hardly dropped below field capacity. Thus, the previous ten days were not enough to dry out due to infiltration and evaporation. This is favourable from the point of drought sensitivity, but it has to be noted that data of deeper sensors are around wilting point (see Table 6.2 on page 251), since early May precipitation did not reach that level. It is supported by the fact that the rain of over 10 mm on 13th May was not able to infiltrate to the sensor at 10 cm. The 70-80 mm precipitation in the following days could also increase soil moisture only later, mostly at 45 cm, where, because of extremely weak permeability (see Table 6.2 on page 251), the infiltrated water was blocked. Significant soil moisture increase at 60-70 cm was caused by the capillary effect of ground water rising from below: water level increased by more than a metre between 1st and 16th May.

With the help of soil moisture changes infiltrated water quantities for both profiles were calculated, omitting the two lowest levels on Tápai-rét: the value was 37 mm in Kelebia, and 23 mm on Tápai-rét. By this calculation it was shown how the refill of water resources – therefore drought sensitivity – depends on the physical characteristics of the soil and compaction due to anthropogenic effects.

In the point of the changes in the amount of precipitation and soil moisture, June can be divided into two parts: until the 21st June it was drier with minimal rainfall, then in the last ten days of the month intensive rainfalls with 55-60 mm were observed, followed by infiltration and drying out. Among the stations Batida and Gencshát were selected for the presentation of June data, and owing to the length of the period daily average values were used. Only values measured at 10, 30 and 75 cm depths were presented for a more clear view of data (Fig. 6.4 on page 254). After the fill-up in May intensive drying out was observed on both locations, soil moisture was under field capacity. It can be seen in case of the soil at Batida, being the most compacted among stations that precipitation in May filled up the topsoil moisture, but lower levels remained dry. In this case the water retained in the upper 30 cm was useful, since it meant reserves for drier periods, but at the same time it could contribute to the formation of inland excess water inundations. The moisture profile of Gencshát is more balanced, but much drier. The only exception here is the depth of 75 cm, where not the meteorological changes, but the capillary effect of the nearby ground water predominated (as a result of rain in May, ground water rose above two metres). Between 22nd and 25th June the increase of soil moisture was high down to 30 cm in Batida, underneath it remained under 2-3 v/v %; while in Gencshát it caused a moisture rise of 5 v/v % at 45 cm, but it was not detectable at 60 cm because of capillary rise. In the last days of June 20-25 mm precipitation infiltrated at both locations.

Conclusions, summary

Based on the data of the half-year-operation of the complex hydro-meteorological stations soil moisture changes of the main soil types of the region could be described. Due to the small amount of precipitation in the second half of 2013 low soil moisture was deterministic in the beginning of 2014, which predicted catastrophic soil drought. The favourable precipitation conditions of the first two months improved the moisture conditions; the sandy soils and chernozems filled up to field capacity by the beginning of the vegetation period (Öreghegy, Kelebia, Kiskundorozsma, Röszke). However, the more clayey (or even compacted) hydromorphic soils were not able to fill up despite of the precipitation, thus they had high water scarcity at early spring. Due to the 50-90 mm precipitation in May and June, the favourable soil moisture conditions for crops remained, infiltration was more than 10 mm. However, compacted soils – including Chernozems – were only able to store precipitation in their upper 10-cm-level, and soil moisture at lower levels was around wilting point even at the end of June (Tápai-rét, Batida, Szentes-Fertő). Special exceptions are the areas where the capillary raise - due to increased groundwater table - was able to compensate the lack of soil moisture due to low infiltration (Gencshát).

The short, only half-year-long monitoring also draws attention on the importance of preserving the original soil structures and preventing compaction, since these processes contribute to the decrease of natural water reservoirs in the Great Plain facing water scarcity (Várallyay et al. 1980).

6.1.2. Satellite based soil moisture estimates for agricultural drought monitoring and prediction

Boudewijn van Leeuwen, Zalán Tobak, Zsuzsanna Ladányi, Viktoria Blanka

Introduction

The continuous estimation of soil moisture allows for dynamic monitoring of the development of the water content in the soil. The trend and direction can be used as early warning for future droughts or inland excess water. Even though soil moisture is an important parameter in many applications, widespread and/or continuous measurements of soil moisture are rare (Patel et al. 2009). Another problem is the limited possibility for upscaling of point measurements to larger areas. These problems can be solved using remotely sensed data. Remote sensing observations cover large areas and can be executed with a high temporal resolution.

Largely two groups of remote sensing based techniques for soil moisture estimation exist. The first uses data from passive microwave instruments. The method is based on the large difference in the dielectric properties of liquid water (~80) and dry soil (<4). The dielectric constant is inversely proportional to the soil emissivity. The soil emissivity can be derived from the microwave satellite data (Owe et al. 1992, Schmugge et al. 2002, Wang 2008). The advantage of method is that it has a solid physical basis and that the data can be collected in all-weather circumstances. Drawbacks are the low resolution of the current passive microwave sensors and the strong disturbance of vegetation to the method (Wang 2008, Vincente-Serrano et al. 2004).

The other group of techniques uses a combination of data collected in the visible, near-infrared and thermal infrared part of the electro-magnetic spectrum. The visible and near-infrared bands are used to derive the relative amount of vegetation, which is often expressed by the Normalized Difference Vegetation Index (NDVI) or the fractional vegetation cover (F_r). The thermal data is used to calculate the land surface temperature (LST). The basic assumption of these methods is that thermal differences in areas with similar vegetation cover are the result of changes in their soil moisture (Vicente-Serrano et al. 2004). Many authors have successfully derived soil moisture estimates based on this principle. For example, Patel et al. (2009) found a strong and significant relationship when comparing surface moisture based on the vegetation-surface temperature space and in-situ measured soil moisture, and Mallick et al. (2009) used ASTER and MODIS data to create LST-NDVI spaces and evaluated their usefulness for soil moisture estimates. They found a fair correlation with microwave sounding measurements from AMSR-A for areas with less vegetation. Advantages of these methods are that they are relatively simple and that the base data is available at global level and at medium spatial and high temporal resolutions.

There are several limitations to the LST-NDVI space based soil moistures estimates. A study area may not cover the full range of vegetation classes (from bare soil to well-developed dense vegetation), and therefore the LST-NDVI space may not be fully determined. Furthermore, since the method is based on remotely sensed LST data only the top few millimetres to 1 cm of soil moisture are "measured", although via the vegetation indirectly also the root moisture is taken into consideration. Also LST and NDVI values derived from satellite imagery may include errors which can propagate into the SMI calculation (Carlson 2007, Mallick et al. 2009).

The aim of this study is to develop a workflow for the dynamic estimation of soil moisture to analyse the water balance at medium scale on the Great Hungarian Plain using MODIS satellite data. Therefore, a fully automatic procedure was developed based on ArcGIS geoprocessing functionality and several Python programs to process MODIS vegetation and land surface temperature satellite data to soil moisture index maps. The Moderate Resolution Imaging Spectroradiometer (MODIS) has 36 bands covering the visible to thermal parts of the electromagnetic spectrum with spatial resolutions ranging from 250 to 1 000 meter. For this study, three Vegetation indices (MOD13) and 36 Land surface temperature and emissivity (MOD11) images were used. Every image covers an area of about 1 100 x 1 100 km.

Methods

The base data for the soil moisture index estimate (SMI) workflow are the Vegetation indices (MOD13) and Land surface temperature and emissivity (MOD11) products. After registration, these can be downloaded free of charge from the USGS Earth Explorer website. For this study, the MOD13A1 product is used which consists of – among others – a Normalized Difference Vegetation Index (NDVI), an Enhanced Vegetation Index (EVI) and a data quality (QA) layer. These data sets are composed of 16 days of measurements in the blue, red and near infrared spectral bands and have a spatial resolution of 500 m. Both indices are computed from atmospherically corrected bi-directional surface reflectances that have been masked for water, clouds, heavy aerosols, and cloud shadows. The QA layer is a binary coded file where for every pixel information is provided about the pixel reliability. The second input data set is the MOD11A1 product, which contains – among others – a layer with daily land surface temperatures (LST) at 1 km resolution. The temperatures are retrieved by the split-window algorithm. This product also comes with a QA layer indicating if the data processing algorithm results were nominal, abnormal, or if other defined conditions were encountered at the pixel level that prevent the product to be used.

The first step of the SMI workflow is importing the NDVI and QA layers from the MOD13A1 product. From the QA layer, all pixels with codes 2112, 2116 or 2120 are extracted and converted to a mask map. Using the mask, only those NDVI pixels are extracted that are of "good quality". The original NDVI data is stored in 16 bits ranging from -2 000 to 10 000, and therefore all masked pixels are multiplied by a scale factor to get the values back to the -1 to 1 range. Finally, a spatial subset of the data set is extracted to match the research area.

In the next step, the NDVI data is normalized to eliminate negative values and create an index between 0 and 1:

$$N=(NDVI-NDVI_{min})/(NDVI_{max}-NDVI_{min}).$$

Gilles et al. (1997) established the relationship $F_r \approx N^2$, where Fr is the so-called fractional vegetation cover. The LST and F_r form a theoretical triangular shaped space, where the wet areas form the lower boundary of the space and the dry areas the upper diagonal (Vincente – Serrano et al. 2004, Carlson 2007, Mallick et al. 2009, Patel et al. 2009) (Fig.6.5 on page 261).

In the next step, the N map is transformed to a F_r map and – by reclassifying the F_r values into equal intervals – 10 sub maps with increasing vegetation thicknesses are created. In the next part of the workflow, the LST map is preprocessed. The original day time land surface temperature values are imported from the MOD11A1 data set and masked based on the QA layer, where all pixels with a value of 0, indicating nominal data are selected as useful data. In many LST files, large areas are excluded, because LST values were not produced due to clouds. The masked LST values are multiplied with the scale factor to receive temperatures in degrees Kelvin. Then, the LST data with a spatial resolution of 1 000 meter is resampled to match the 500 meter resolution and geometry of the NDVI data. Finally, a spatial subset covering the research area is created from the LST data.

The presented method assumes a linear relationship between the LST and the soil moisture within one F_r class, therefore for every F_r map the pixel with the lowest and highest values are extracted from the LST map. The pixel with the lowest LST value in a particular F_r class gets a soil moisture value of 1 and the pixel with the highest LST gets a soil moisture value of 0. The soil moisture value SMI for F_r map i for the intermediate pixels was calculated by:

$$SMIi= (LST_{min}-LST)/(LST_{max}-LST_{min})+1$$

This results in a SMI map for each of the 10 F_r maps. In the final step, all SMI_i maps are combined to form the SMI map for the total study area (Fig. 6.6 on page 263). This map shows the spatial distribution of soil moisture in the area at a particular moment in time, where the 0 indicates the lowest soil moisture and 1 the highest soil moisture.

Soil moisture estimates using satellite data

SMI maps were calculated for a period from 1st March 2014 to 13th April 2014 to determine the change in the spatial distribution of soil moisture in the area. The resulting maps for four days are shown in Fig. 6.7 (page 264). The rivers in the south and the forested areas are clearly visible in the images. Also, the total study area shows a larger variation in soil moisture on the first and last day than during the intermediate days. LST data can only be collected from cloud free areas, therefore on 31st March and especially on 3rd April in many areas data is missing and at those locations the SMI can not be calculated.

Following similar approaches by Wang (2008) and Mallick et al. (2009), satellite based SMI values were compared to ground soil moisture (SM) measurements. At both sides of the border eight stations were established and since end of January 2014, every hour soil moisture data at 6 different depths (from -10 to -75 cm) is collected using Decagon EC 5 volumetric water content sensors (Jovanović et al., 2013).

At the locations of the measurement stations, soil moisture index values were extracted from the satellite based maps. Data from two stations were omitted because they are not representative for the neighbouring 1 km zone (the spatial distribution of SMI mapping). Figure 6.8 (page 264) shows graphs of SM ground measurements and the satellite derived SMI values for two different days. The relationship between the two data sets during the period when ground measurements are available is not very strong. The main reason for this is the large difference in scale between the data sets. The spatial representativeness of the point measurements is limited to maximum several hundreds of square meters around the measurement stations, while the satellite data is an integrated measurement of a 1 km^2 area. Furthermore, the ground data is measured at depths of 10 cm (Hungary) and 20 cm (in Serbia) while the satellite acquires only surface temperature. Figure 6.9 (page 265) shows the precipitation and SM in the observed period at two Hungarian stations. The first part of March (before the 13th March SMI map) was a long dry period, and higher correlation (R^2=0.57) between satellite based SMI and ground station based SM is observed on 13th March. Hardly any connection between parameters was detected (R^2=0.14) on the 22nd March (after minor precipitation), which can be due to different water holding capacity of the soils and due to infiltration. Again, higher correlation (R^2=0.54) was detected on 31st March which was after a rainfall event.

Conclusions and discussion

SMI maps based on MOD11 and MOD13 products can be created automatically with the presented workflow. The maps provide a good impression of the spatial distribution of SM at a specific moment in time. The fractional vegetation maps can be enhanced by incorporation of soil type information in the bare soil class.

The relationship with the local point measurements is not very strong, and therefore absolute calibration of the relative SMI maps is not feasible. Using longer data series can improve the calibration and the effects of rainfall events on correlation can be investigated. Many possible improvements to the method exist. Among others, absolute calibration of the SMI values can improve the relationship with the point measurements. Also, additional information on the soil type may improve the creation of the LST – F_r space. The use EVI or LAI instead of NDVI may result in better F_r data and therefore improving the LST – F_r space as well.

6.2. Potentials of surface water monitoring and modelling

6.2.1. Monitoring potentials of long-term changes of surface water cover

Ferenc Kovács

Introduction

On the Great Hungarian Plain, natural in almost 75% in the second half of the 18th century, the extent of water covered areas had significantly decreased by the 1960s due to flood and inland excess water regulations (Somogyi 2000). Today the rate of permanent or temporary water cover is slightly more than 2-3%, although it was 30-35% prior to water regulations. As a consequence of water regulations, the regular water cover on landscapes along rivers ceased to exist from the end of the 19th century onwards, and then people further decreased the size of areas temporarily or permanently covered with water by inland excess water regulations. Besides human impact, the intensifying aridification of the climate has further aggravated the conditions of areas depending on the amount of precipitation since the 1980s (Iványosi 1994, Boross and Biró 1999, Hoyk 2006). The extent of (secondary) saline areas, results of human activity, can further increase even today, as the warming and drying climate contributes to the shioft of natural soil processes towards salination and steppe formation (Rakonczai and Kovács 2006, Csorba 2011).

Water cover as a local feature may be the most dominant landscape factor (indicator) from a geographical point of view, whose dynamics is of high importance in the accelerating landscape degradation processes. It is important to provide information about spatial and temporal intensity of changes, which can be examined more precisely and in more details due to the developing possibilities in geoinformatics.

Methods

Spatial data (starting from the first available map, furthermore occasionally available topographic maps, satellite images) were processed jointly to describe the development of the expansion of wet areas, which is a determining factor of landscape formation. The spatial differences and the speed of the observed processes have great importance. In case of the satellite images, in order to make a comparison, the most favourable (wettest) conditions were examined every year on the study area. If possible, images taken in June were included into the dataset. The wetland condition was regarded critical, if even this most favourable status shows bad conditions.

Besides analysing the long-term changes, it is important to know the rate of variability, which may affect the opinion on the changes. The more variability a patch features, the more uncertain the trend of the change is. Analysis in higher temporal and spatial resolution will variability assessments possible (Kovács 2009). Extreme conditions also need to be evaluated in order to analyse the changes of wetlands. Local effects of climate change are the increasing frequency of precipitation falling in a short time and the increasing drought frequency. The fast fill-up of lake beds and the possibility of fast and permanently drying out also need to be considered. According to the method of temporal analogy, a study of an extreme period with high temporal resolution can be applied as a good reference. The processes occurred in the extreme year 2000 could be typical regarding climate change in the near future. That explains why variability was mapped for this period. 22 satellite images are available for the period between July 1999 and October 2003 for the study area located in the Danube-Tisza Interfluve (Fig. 6.10 on page 269). The investigation is interesting because the impact of a shorter period with humid years can be analysed within a longer, unfavourable (becoming arid) period of time.

Water content in the infrared range of multispectral images can be well delineated, so we apply automatic classification, where the given 30 classes were analysed visually.

The moisture conditions were determined by wetness index:

$WI_{ETM+} = 0{,}263_{ETM1} + 0{,}214_{ETM2} + 0{,}093_{ETM3} + 0{,}066_{ETM4} - 0{,}763_{ETM5} - 0{,}539_{ETM7}$

where: ETM1...ETM7: different wave length ranges

Vegetation cover was defined by normalized vegetation index:

NDVI = (ETM4–TM3) / (ETM4+TM3)

Maps with classifications of "open water cover", "area with high water content", "wetland", and "dry surface" were created based on complex queries, mostly considering the automatic classification and WI index photos. On the topographic maps the legend was used for identification while digitalizing these classes.

The spatial appearance of aridification was evaluated in terms of variability. Certain patches, compared to the reference conditions typical until 1962, were difficult to be classified in a long process (e.g. the surface was once covered by water, than wet and dry). The "constantly wet" areas had always been water-covered or wetland patches. Patches as per average years can be assessed as 'in general covered with water', while areas flooded by high waters were ranked in the category of "impermanenty covered with water" which will belong to the "probably not drying" category while assessing aridity. The "moderately becoming arid" category refers to still water which became swamps and the dried out former swamps, and waters appearing only at large floods also belong to it. Patches of old wetland areas remaining dry in general are a part of the class of 'becoming arid'. The area of 'becoming seriously arid' was previously covered with water until the 80s, but not since then. An optimistic and a pessimistic scenario were outlined at the assessment of degradation process. In the optimistic scenario, at the questionable patches, the more favourable (wetter) conditions were considered, while in the case of pessimistic scenario drier conditions were taken into consideration.

Results

It is visible from the data showing the expansion of water-covered areas and wetlands (Fig. 6.11 on page 271) how difficult it is to recognize the process of changes owing to the variability of the area of 13,000 hectares. There are differences between the years, but even a shorter favourable period can set the old situation back "out of the blue".

The most prominent change could be seen 100 years after the 1880s, when 84% of the water-covered areas disappeared. In 2010 the almost natural condition was activated in hydro-geographical point of view. Watery conditions similar to the reference status appeared due to the climate change, although there was a need for extreme precipitation.

In the years following the huge inland excess water inundations of 1999-2000 and that of 2006, low levels of inundation values were experienced, similar to the 1980s. As there was no sufficient water resupply, ¾ of the water had disappeared by 2001, within two years, and has remained so permanently. There is a striking difference between 2006 and 2007, when about 50% of open water and 85% of swamps disappeared within a year. It is evident that the impact of some more favourable years is not sufficient to stop the unfavourable processes going on since the 1970s.

Based on our definition regarding the datasets of nearly 130 years, the pessimistic approach says that 33.5% of the areas will become arid, while the optimistic approach says it is 6.5%. The worse scenario projects that 6.3% of our areas is getting seriously arid. In the case of a change analysis, those areas are important which are considered stable from the point of view of change assessment. Changes on versatile areas are more difficult to register, and a process can be more dangerous if it threatens the more stable phenomena, too. Results of the pessimistic and optimistic viewpoints were refined by the spatial results of variability in the long-term analysis. Only results of patches with little variability were further evaluated (Fig. 6.12 on page 272) According to the more precise map, the value of aridity for the pessimistic approach decreased to 24.7%, the value of the optimistic approach was reduced to 5.6%. Even the most advantageous condition projects problems in the south-western, south-eastern and eastern parts of the area and in the Zab-szék area. Observing the relationship between precipitation and hydro-geography, and the geographical processes of the area, the pessimistic approach is more likely to be realistic.

6.2.2. Monitoring spatial and temporal appearance of temporary surface water covers (inland excess water)

Zalán Tobak, József Szatmári, Boudewijn van Leeuwen, Mesaroš Minučer, László Mucsi

Introduction

As it could be seen previously, both periods with water shortage and with inland excess water occur frequently owing to the fluctuating water supply on the study area, therefore water management and the planning of water retention are of vital importance. It is important to monitor inland excess water in order to plan water management and water retention. When monitoring inland excess water inundation, the knowledge and continuous follow-up on the spatial and temporal extension of the affected area provide significant information for the understanding of the procedures of development and disappearance. It also contributes to the better reliability of the forecast. Inland excess water and/or saturated soil layers on areas under agricultural cultivation are harmful to the vegetation in the short run, while permanent water cover could have positive ecological impact on nature-related (for example meadow, pasture land) areas. As a consequence of climate change, extremes could be observed in consecutive years, or it may result in periods with extremely high precipitation (resulting in inland excess water) and dry periods (resulting in drought) within the same year in Hungary. The water surplus of years with high precipitation could be applied to mitigate water shortage by appropriate water management (water diversion, water storage). This could reduce the damages caused by inland excess water. All these draw the attention to the necessity of mapping the inland excess water inundation, which includes the monitoring of its development, size, durability, frequency and disappearance.

The methods used to study the spatial and temporal pattern of inland excess water can be divided into two large groups: (1) observations and assessments based on field (Fig. 6.13 on page 274) or remote sensing measurements and (2) calculations based on the factors influencing its development on the basis of experimental weighting factors (van Leeuwen 2012). In the latter case, models with several input parameters will result in hazard maps.

In situ field surveys based on topographic maps with the scale of 1:10000 or 1:25000 – and on the knowledge of experts about the local conditions – are subjective (for example to indicate the "border" of inland excess water patches). The frequency maps created from them do not correlate with one of the most important factors of inland excess water formation, the surface relief (van Leeuwen 2012). Furthermore, surveys take a long time. However, by these surveys, inundation data are available retroactively for decades, which can be used to calibrate experimental models with restrictions (for example, at an appropriately small scale).

It is possible to record terrain data by high precision GPS devices. However, the marking and approaching the borders of inland excess water and saturated soils may be problematic. Moreover, it is more time-consuming than the above mentioned data collection on topographic maps from the roads encompassing the patches.

Geoinformatic possibilities of mapping inland excess water

Through the development of remote sensing technologies – such as data collection and data processing - not only field assessments, but mappings based on the evaluation of aerial photos and satellite images (Rakonczai et al. 2001) are also becoming more focused on. As spectral features of water and water-saturated soil surfaces are typical (nearly complete absorption above infrared range), they may also be defined by automatic procedures. In the near infrared range (900-1200 nm), most of the satellites observing the Earth make images, so the use of these data seems obvious. However, the appropriate spatial and temporal resolutions are limiting factors. As the real range for the inland excess water patches is from some 10 to 100 metres, the 30-meter resolution of Landsat (E)TM(+) is still acceptable for this purpose. Inland excess water may develop and disappear relatively fast, so the temporal resolution (the time elapsed between the images taken of the given area) of the utilised satellite images need to be as high as possible. The 16-day recurring cycle of Landsat, which under unfavourable circumstances may coincide with cloud cover, is an important limiting factor. RapidEye satellite images could be the optimal solution, which are able to provide data even daily, at relatively affordable prices.

The assessment of multispectral images, in this case the delineation of areas affected by inland excess water, could be performed by manual and (half) automatic procedures (Fig. 6.14 on page 274). In the first case the precision of the inundation maps depends on the spatial resolution and on the (partially subjective) decision of the specialist conducting the evaluation. The method is time-consuming but faster than a field survey, and can cover larger areas at the same time. Unsupervised methods (clustering) divide the spectral (feature) space into a predefined number of clusters and the pixels are labelled based on their location. The meaning of the clusters (for example, inland excess water, vegetation, dry soil) are need to be defined by the experts. Although this is a fast procedure, the results of separate images are difficult to compare. In the so-called supervised classification, classification is preceded by the training phase, where the spectral features of a land cover to be mapped are recorded. Afterwards the algorithm will automatically label the pixels based on the similarity.

As inland excess water patches do not have sharp borders (fuzzy borders), it may be necessary to analyse their ratio within a pixel element. Spectral mixture analysis provide this information, and can efficiently separate water surfaces from their environment. Another novel method of classifications is the application of artificial neural network (van Leeuwen 2012).

Satellite images provide data in same quality at relatively low cost with wide area coverage. As for the spatial and temporal resolutions, which are vital for the mapping of inland excess water, they cannot compete with aerial photographs. The spatial resolution can be arbitrarily adjusted by the flight height (0.1-1 m) in case of aerial photos, and the time of imaging depends only on weather conditions. A disadvantage is that imaging is frequently made only in the visible spectrum (RGB). The same methods may be applied for the assessment of aerial photographs as for the satellite images, but the inland excess water patches can be delineated geometrically in more detail and more precisely due to the larger amount of geometrical information (Fig. 6.15 and Fig. 6.16 on page 277-278).

Possibilities of complex inland excess water mapping

In the traditional classification methods only data layers, whose histogram shows natural (Gaussian) distribution, can be used. Remote sensing images or digital surface models meet this criterium. In order to include the effects of human activity and artificial structures, more advanced methods are needed. There are no limitations to the classification carried out with artificial neural networks, thus the delineation of the current inland excess water inundations can be achieved by considering many factors (data layers) with more reliability (van Leeuwen 2012).

Model-based approaches will result in hazard maps. The Pálfai-model mainly considers the natural factors from the natural and anthropogenic affecting factors of inland excess water formation. These are the hydrometerological parameters, the hydraulic conductivity of the soil, the geological parameters (depth and thickness of impermeable layer), the changes of the depth of groundwater level in past decades, the elevation (relief) and land use. Human activity may be present in the system through the characteristics of soil, relief and land use.

6.2.3. Assessment of different hydrological modelling software on a lowland minor catchment

Balázs Benyhe, Tamás Právecz, György Sipos

Introduction

Hydrological models were developed for understanding and quantifying the factors of the complex hydrological cycle by mathematic, physical or empirical functions on a well defined hydrological system or catchment (Singh and Frevert 2001). On hydrologically extreme areas, such as the lowland small catchments of the Carpathian Basin, more accurate description and forecast of the water balance is important objective, since only a few exact data are available about evaporation, runoff, infiltration and water storage conditions of the area. During this research, water balance (e.g. volume of runoff from the catchment, temporal variation in runoff, volume of water storage from infiltration) was defined on the catchment of the Fehértó-majsai Major canal by different hydrological models. The results of the modelling can support the realization of water management and planning projects in the drought prone sand land region, where only a few objective data could support the planning.

Study area

The modelling was carried out on the catchment of the Fehértó-majsai Major canal. The catchment is located mainly on the Dorozsmai-majsai sand land and partly on the South-Tisza Valley region (Dövényi 2010). The main waters on the catchment are the major canal, its 6 tributary canals and 1 main lake (Fig. 6.17 on page 280). The canal density is 0.68 km/km^2 on the basis of the total length of canals managed by water directorate. The area of the catchment is 305 km^2, however the analysis was carried out on the area upstream from the Szatymaz discharge recording cross-section (approximately 290 km^2). Below the Szatymaz cross-section, the further 4 km length section of the canal drains the water to the Fehértói major canal. Low slope conditions exist on the catchment, despite the ridge-like character of the area. The slope of the major canal is 0.78-1.16 m/km on the upper reach and 0.27-0.78 m/km on the lower reach. The annual precipitation is only slightly above 500 mm in the area. The dominant soil types are humic sand of low fertility and blown sand, nevertheless high proportion of the area is agricultural land (mainly small-parcel arable land). Farmlands also situated on large areas on the catchment. Natural or semi-natural areas (sandy grasslands and wetlands) are located mainly on the Homokhátság Region.

Applied input and control data for modelling

For analysing the relief and the topographically determined runoff directions on the catchment, digital elevation model (DEM) of 5 m resolution was used. Average precipitation for the catchment was calculated on the basis of 4 meteorological stations and this average data was used as input for the models. For defining soil conditions in the model AGROTOPO digital soil map was employed. Accuracy of the modelling was improved by including spatial distribution of soil water household. Land use data was derived from CORINE (CLC50) and from MADOP ortophoto database. Control data was gained from the Szatymaz discharge recording cross-section, where a flume and automatic stage recorder are located. Discharge data was calculated for the whole studied period, based on regularly updated stage-discharge function.

Assessed models and the modelling results

Modified Budyko model

The original Budyko model calculated the mean annual runoff from precipitation, evaporative heat transfer and annual radiative balance data (Nováky 1985). In the modified model only annual precipitation and mean temperature are included because of the difficulties in calculation of annual radiative balance (Keve and Nováky 2009).

The Budyko model in this form is not suitable for modelling at shorter time steps (e.g. monthly or daily), thus calculations were performed at yearly and decadal time steps. The mean annual runoff was obtained from the model in mm/year. Both the spatially concentrated and spatially distributed variants of the model were assessed in the research. The spatially distributed model was built up in ArcMap software, while spatially concentrated variant was built up in Microsoft Excel software.

Based on decadal time step modelling, the mean annual runoff is 14 mm/year, while the control discharge was 16 mm/year for the whole catchment. This model result with its 10 % error is acceptable; however modelling at yearly time steps produced more inaccurate result.

The modelling at yearly time steps was carried out only with the spatially concentrated model variant, based on results from decadal analysis. The results calculated on the basis of annual mean data show much higher deviations from the observed data (Fig.6.18 on page 282).

HEC-HMS modelling software

The spatial data for modelling in HEC-HMS software was generated by the HEC-GEOHMS toolbar in ArcMap software. The model was run with two variants. In both variants, discharge time series were simulated at the outlet point (Szatymaz cross-section) of the catchment. In the first variant keeping the conceptual character of the model was in the focus, meaning all possible input parameters were tried to fit into the model. However, in this way hydrological processes could be only broadly quantified. Modelling hydrological scenarios, preparing forecasts or calculating hydrological parameters could be achieved with the physical based model variant only by measuring and calculating the necessary input parameters.

In the second model variant all those model parameters were removed, which had no effect on the modelling result in the first variant. The most important modification was the integration of the precipitation, infiltration and evaporation data. The precipitation data were corrected based on the result of McCuen (2005) and the model was run with this corrected precipitation data and a hypothetic 100 % runoff coefficient. This second model variant can be considered as a transition between empirical and conceptual models.

The modelling results can be calibrated with the time of flow concentration and the storage time, thus good accuracy can be reached in case of the magnitude of the discharge, however the shape of discharge curves is affected by modelling errors. From the long time series, modelled and measured discharge curves of two years are shown as examples on Fig 6.19 (page 284).

The modelling results of HEC-HMS indicate that greater accuracy could be reached by the simplified model (based on reference values published in scientific literature) than the physical based model variant. The main reason of this is the data demand of the model. This physical based model requires input data, which are not available and the modelling with estimated values of these parameters led to false results.

On the basis of the modelled discharge curves, the HEC-HMS software overestimates the runoff on plain areas. The modelling software is sensitive to the precipitation amount; however in low relief areas (slope 1-2 m/km) the precipitation has less relevance on discharge than the water household capacity of the surface or the temperature.

MIKE modelling software

From the wide range of MIKE software products, MIKE 11 and MIKE SHE were used. MIKE 11 is a one dimension (1D) river and channel modelling software, while MIKE SHE is 2D integrated catchment modelling software. MIKE SHE is suitable for physical based modelling of all hydrological sub-processes and it can model the interactions between the elements of the hydrological system. The two modelling environments were joined, thus the interactions between the water flow and the catchment could also be interpreted.

Based on the results, the model simulate the discharge well in less humid periods, the measured and simulated time series of discharge show good similarity. However, in more humid periods the model overestimates the rate of runoff; as it was experienced in case of HEC-HMS (Fig. 6.20 on page 286). The higher simulated discharges are caused by the soil saturation and the accompanying rise of groundwater level in the model, which causes higher subsurface inflow than the real (Fig. 6.21 on page 286). Therefore the more accurate determination of the hydraulic parameters of the soil and hydrodynamic properties of the groundwater is essential to improve the model accuracy.

Summary

The results of the modelling show that the modified Budyko and HEC-HMS have limited efficiency in simulating runoff conditions in case of small lowland catchments. The main reasons are that the models put little weight to some important hydrological sub-processes, e.g. the subsurface water movement, which is an evidently significant process on catchment with low or no runoff.

The Budyko model excludes all input parameters except precipitation and temperature to facilitate easier application. Therefore all features of the catchment are described by quasi-constant values, which caused inaccurate simulation results. Consequently, Budyko model can be suitable only for simulating long-term changes of large catchments.

The HEC-HMS model makes possible more detailed analyses, however it is quite difficult to build up a hydrologically correct physical model. In the tested model variants, the infiltrated water filled up the available pore space or evaporated or it is leaked from the system by constant value. The subsurface water movement cannot be modelled, which was resulted in inaccurate runoff values. The semi-empirical model variant of the HEC-HMS provided more accurate simulation results, than the physical based variant, because the type and quantity of input data are not appropriate for building up a proper physical based model. The semi-empirical model variant can be suitable for modelling the runoff in rainy periods, however by the exclusion of subsurface water movement and water storage the model is not appropriate for hydrological modelling on lowland catchments.

Based on the results, the necessary input data for MIKE integrated hydrological modelling software can be defined more objectively and in more detailed. The built-in dynamic feedback processes of the software make possible to create a physical base model, by which all subjective parameters can be eliminated. Moreover the software makes possible to check the spatial and temporal changes of all sub-processes. This is a great progress compared to the other assessed models. However, it should be noted also in case of MIKE that more accurate modelling results could be achieved by including measured data of some parameters (e.g. hydraulic conductivity of soils). Thus, future research in the analysed catchment should focus on this problem.

6.3. Potentials of groundwater monitoring

József Szatmári, Károly Barta, Zoltán Csépe, Zsolt Fehér, Brkic Miodrag, Djordje Obradović, Zorica Dudarin, Vasa Radonić

Introduction

Groundwater observation came to the fore in Hungary in the 19th century because of agricultural considerations, to create appropriate conditions to grow rice. The first irrigation model farms were established in 1863, in Sárrét, Fejér County, then later, in 1878, in Pékla-puszta, on the Great Hungarian Plain. Though the knowledge of the hydrology of waterways supplying these sites was essential, attention turned to groundwater observation also at that time. The first observations were done in 1866; a five-year data set was created afterwards, between 1876 and 1880, by regular measurements in the region of Szeged and Debrecen (Szalai 2003). The observation network gradually expanded, and in 1929, it consisted of 149 wells with a density of 80 km, and specifically served the observation of groundwater. The national groundwater level observation network started to be built in 1933. Regular measurements were carried out in the network, and typical monthly and yearly water levels were defined. The network reached its largest extension in the 1970s with 1500-1700 wells.

The technology of detecting groundwater level has changed a lot compared to its beginnings, and the development of digital measuring devices and remote sensing stations meant significant progress. Computer technology was also a step forward in this area. The automatic instruments were able to carry out measurements, and some of them could also forward data. With the appearance of wireless technology, physical contact with the measuring instrument to download data has become unnecessary, and data forwarding can be carried out via a GSM network, as well. One of the modern instruments, the system presented here (the network was built by the University of Szeged and the University of Novi Sad, in the framework of the

MERIEXWA project), uses ultrasound technology to measure groundwater levels (Fig. 6.22 on page 290). Data storage improved continuously, together with data collection methods. The appearance of different GIS softwares has simplified not only data storage, but also data processing. These programs allow the placement of processed data sets on the base map digitally, instead of the former offset printing technology (Barton et al., 2013).

Study area

The main selection criterion for the project area was the inundation hazard. This cross-border problem affects both the left and the right banks side areasof the Tisza River, in Hungary and in Vojvodina, as well. Thus, the south-eastern part of the Danube-Tisza Interfluve blown sand area and its cross-border extension, the Marosszög, lying north of the River Maros; Torontal, on the southern side of the Maros; and the territory of Banat in Vojvodina were chosen. The causes of the appearance of inland excess water are extremely diverse in these areas, which are different in their topographical, geological and soil features. The very high clay content is responsible for the formation of inland excess water on the alluviums along the Tisza and the Maros (Marosszög, Torontal, Banat). Inland excess water can be found here both from groundwater sources and accumulation. Regarding the physical type of sediment, we can find sandy and sandy loam sediments mainly to the west of the Tisza River, and the water flowing out from under the blown sand ridge is responsible for inland excess water here. When defining the western boundary of the project area, it was important that not only the areas of inundation hazard could be monitored, but also the higher elevated feeding areas where the rising groundwater levels can indicate the inland excess water situation arising. Thus, the appearance of inland excess water around Ásotthalom and Kissor is less likely, in fact, the area faces serious water shortage, except for extremely wet years (e.g. 2010). Together with the monitoring network in Vojvodina, here we also have the possibility to detect depression due to water abstraction at Subotica.

Methods

Defining the location of well network

In groundwater modelling, a limited knowledge about the features of the test environment can be gained, since the object of research is almost never entirely known (Bárdossy et al. 2002). A cardinal factor of the reliability of groundwater modelling is the establishment of a representative monitoring network. Constructing the monitoring network was realized in the framework of the MERIEXWA (HUSRB / 1202/121/087) Serbian-Hungarian project in 2012 and 2013.

In geostatistical meaning, groundwater monitoring is considered to be reliable if the information from the sampling strategy realistically characterizes the groundwater level (Bárdossy et al. 2002). Consequently, the deployment strategy of the well network, and the density of measurement intervals are essential determining factors of estimate reliability (Geiger 2007). When designing the sampling strategy, (1) the size, shape, and spatial location of the individual geological formations (2) the spatial distribution and variability, the effective range and anisotropy of their features, and (3) the effects of other geological processes, structures and influencing environmental factors must be taken into account (Füst and Geiger 2011). The location of the 20-25 wells on both sides of the border was planned in a way that they form a joint groundwater and inland excess water forecast system with the existing wells on the national-level groundwater monitoring area.

University-developed sensor network

The water stage recorder consists of two main components. One of them is a PROTON mote, which stores the data, and realizes communication with the computer. The other is an acoustic sensor card. The two-part metering device emits a sound signal that passes down the tube until it reaches the level of the groundwater, from where it is reflected, and gets back to the microphone, and then the time having elapsed form the time of emitting the sound is recorded. The sensor emits a one-millisecond sound (click) through the speaker for each measurement, and then makes a 16666 Hz audio record. The echo return time is measured by a simple digital filter using the audio record (we know the frequency of the sound source, and we look for a rapidly rising signal strength). The digital filter looks for three reflection times (range), and it assigns the number characterizing the growth of sound strength detectable at each (score). The results of the measurement are stored in the internal flash memory of the device.

Aspects of processing geostatistical data

The reliability of measurements and observations in the project area can be affected by human and technical factors (intermittent irrigation, pumping) on small scale (Rakonczai 2011), since they can distort data series of groundwater monitoring stations. Measurement errors can also be caused by failures of the sensors used. In the spatial modelling all these can result in showing a depression larger than the existing one. In our case, the errors mentioned are easily recognizable in the timeline of groundwater data due to the measurements continuously performed (repeated sampling). The errors can be filtered by simple statistical methods. General geographical conclusions can be drawn from the mathematical nature of the data sets of individual wells related to space and time (outliers, distribution analyses, involving auxiliary information on interpolation, and classification).

When considering the reliability of the spatial expansion of information, it should be taken into account that the geological environment is not homogeneous in space, and therefore the spatial variability of hydro-geological parameters seriously affects groundwater flow conditions. Though the aim is an accurate and deterministic description of the project area, it is nearly implausible in practice due to the limitations of knowability (Caers 2005).

Creating a proper model may be the most problematic task in hydrological modelling. Groundwater level is a random variable in the geostatistical sense (Delhomme 1978). The separate observations correlate along a certain spatial regularity (de Marsily 1986). Unlike any other method, kriging interpolation is suitable to involve this spatial structure through the variogram models of a given momentum of data sets. Kriging is especially suitable for estimating the changes of groundwater resorce, because it gives the best local, linear estimation of the groundwater level by minimizing the variance of point errors. However, the reliability of kriging (kriging weights) is significantly influenced by the geometrical structure of data points (Geiger 2012).

In groundwater modelling, instead of minimizing the estimation error in measuring stations, the aim is to quantify the reliable estimates of the unknown grid points, and the uncertainty of the estimates. The Gaussian stochastic simulation realizes a large number of alternative, but equally likely estimates, using information of data points with a dynamically changing geometrical structure. The changing geometry means the continuous inclusion of the previously estimated grid point values into the next grid point kriging, using a simple random traversal path (Caers 2005). On one hand, the estimate confidence intervals (reliability for a given level of significance), on the other hand, the expected value of groundwater level can be determined from the distribution of the estimated values for each grid point of a high number of stochastic images.

Based on previous studies (Pálfai 1994, Rakonczai 2011), a significant relationship can be assumed between groundwater and topography. Thus, digital elevation models in spatial extensions significantly improve estimation results. Co-kriging, more exactly, the Markov 2 model was used from a large number of geostatistical solutions, because it facilitates the involvement of regularly sampled secondary data with high information content (Journel 1999). To carry out co-kriging (1) groundwater modelling, (2) DEM and groundwater modelling, and (3) DEM variogram modelling are needed. As groundwater is constantly changing in time, models (1) and (2) must be set up for each momentum when modelling groundwater. In contrast to the spatial location of groundwater wells, the information content of the DEM is of uniform coverage, its semi-variogram is considered to be much more reliable (more representative), and also constant for a longer period of time. The variogram model of groundwater and the semi-variogram model of the groundwater-DEM cross analytically follow from the DEM and from groundwater data when carrying out collocated co-kriging with the Markov 2 model (Journel 1999). The NEWCOKB3D algorithm of the GsLib function package can be applied during co-kriging, and the modified version of SGSIM_FC during sequential co-simulation.

1. The estimation of a given momentum of groundwater comprises the following steps: Statistical analysis of data error (filtering outliers), estimates of incomplete and erroneous data
2. Establishing the parameters necessary for the Markov 2 model for the separate momentums:
   - linear regression relationship between the DEM and groundwater levels
   - the variance of groundwater
   - determining the correlation coefficient between the DEM and groundwater
3. Modelling groundwater as well as the cross-semi-variogram with the application of the Markov 2 model
4. Generation of GsLib parameter files
5. Smoothing groundwater data histograms, normal score transformation of the smoothed data
6. The realization of a large number of groundwater levels in the case of stochastic co-simulation, executing the interpolation in the case of kriging
7. Normal score re-transformation of data obtained
8. Post-processing of the data
   - Expected-value type estimate
   - Determination of confidence intervals

Results

After the extremely dry years of 2011 and 2012, the groundwater level dropped below 3-3.5 m even in the areas most vulnerable to inland excess water (Torontal, Marosszög), and it became constant at under 5 m in the higher parts of the blown sand area (e.g. around Ásotthalom). The extremely wet first quarter of 2013 unexpectedly changed the groundwater table: 1.5 to 2 m water level rise occurred in the vast majority of wells in a little more than three weeks, between 15th March and 10th April (Fig. 6.23a on page 297). One of the preconditions of this was the high precipitation; the Batida measuring station of ATI-VIZIG recorded 43 mm precipitation in January, 52 mm in February, and 108(!) mm in March. The shallow groundwater table and the geological structure also played a significant role in the rapid water level rise.

The change was not so pronounced in the case of a few wells; the spring water level rise happened at a smaller scale, its maximum was a few tens of cm. The reasons for this are that the effects of precipitation occur later and are mitigated when the groundwater levels are lower, in addition, the morphological situation and the near-surface sediment permeability are different. Due to these factors, a similar curve of the recorded values was experienced in the cases of wells with very different hydrological features (Fig. 6.23b on page 297).

The well in Ásotthalom, shown in Figure 6.23b (page 297), deepened into a highly permeable sandy sequence, but the effects of precipitation were barely noticeable because of the low water table and significant lateral flow-off. Interestingly, the operation curve of Marosszög well no. 23, situated in a low elevated alluvial sediment area with a mixed matrix of clay-silt-sand, was the most similar.

In the latter case, the near-surface layers of low permeability have a decisive role: the designation of the well was carried out according to the Perger-plan, and the well is situated in an area more clayey than the surroundings based on the shallow subsurface geological map. This is also supported by our field observations; inland excess water spots in arable land around the well were observed at the end of April (Fig. 6.24 on page 297). This show a good example for accumulative inland water here, as the water level of the well was still at nearly 4 m deep from the surface under the inland excess water spots.

The same well highlights the importance of the optimization of the well network, as well as the importance of spatial modelling. Fig. 6.25 (page 298) shows the absolute water levels of well no. 23 and the surrounding three wells. At the beginning of the recharge period the water levels of the previously discussed two wells were at a much lower elevation, and, despite the similar relative water table, they had gained about 1-1.5 m "advantage" in a matter of weeks, creating a specific, large groundwater depression in the environment of well no. 23.

Experiences and other opportunities of the monitoring system

During the period of 1-1.5 years of its operation, the groundwater monitoring network and the complete Hungarian-Serbian system modelling groundwater levels were not tested in extreme weather situations, or in the subsequent periods of inland excess water or drought. However, the previously described results of a few wells show that the new measurement, data transmission, and data processing system meet the requirements. Further tasks are the integration of the university-run network into the regional systems operated by the Hungarian and Serbian water management directorates, and a long-term, reliable, and efficient operation of a joint monitoring system, with which we can substantially contribute to the scientific assessment of the impacts of climate change in the Southern Plains and North Vojvodina.

6.4. Potentials of vegetation monitoring

Ferenc Kovács

Introduction

The Danube-Tisza Interfluve is basically sensitive to changes, the extremities are intensified here by the sandy surface, apart from the water shortage due to the climate. According to previous analysis, the landscape ecological value is expected to decrease in the case of the forests in the Danube-Tisza Interfluve (Mezősi et al. 1996). Based on the national climate scenario, biomass will be lost on 80% of the forests (Lasch et al. 2002). Among the ecosystems of the Great Hungarian Plain, the forest is the most balanced and the most natural habitats in spite of the fact that maximum of 20-25% of the forests in the Great Hungarian Plainare natural or semi-natural forests (Járó 2000). They can preserve moisture effectively; therefore they are able to indicate also the occurrence of permanent aridity.

Several types of vegetation indices are applied to monitor vegetation. The European Drought Observatory (EDO) assesses fAPAR anomaly so as to provide up-to-date information on the status of vegetation with remote sensing. The datasets of NDVI and EVI indices in MODIS sensor are of high temporal resolution. Many publications prove that they can be applied for a wide range of purposes such as the analysis of wind erosion (Mezősi et al. 2013), changes of wetlands (Kovács 2007) and drought sensitivity (Ladányi et al. 2011), which have been carried out in Hungary. Various phenological changes (drought, pest damages, clear cutting) can be indicated from the calculated parameters of the annual curves (area below the curve, slope, shape of the curve, amplitude) drawn from vegetation index data (Fig. 6.26 on page 300).

Methods

The sensitivity of the planted forests to water shortage in a study area in the Danube-Tisza Interfluve was examined on the basis of MODIS NDVI index datasets. The aim was to monitor the changes of natural water balance in the forests of the flat area of the Danube-Tisza Interfluve for the summer half-year for the period of 2000-2011. Only those pixels of the small geometrical resolution photos were taken into consideration, which were covered at least in 60-65% by forests. On the study area the analysis of only three types of surface cover were carried out, because the area is characterized by low forest cover (14%), and heterogeneous land use: the categories of deciduous, coniferous and mixed forest.

The forests were delineated and the vegetation dynamics were assessed on the basis of CLC50 maps and 16-day-composite images of the 250-meter resolution MODIS vegetation index images (NDVI, EVI indices), which can be downloaded from USGS Data Pool database. The stress effect due to water shortage can be evaluated on large areas with high spatial resolution by the evaluation of satellite image-based indices. Not only the assessment of average, extreme and total values for the 12-year period was carried out based on the 16-day-composite images of NDVI and EVI , but also so-called 'average photos' were taken of certain surface-cover classes. Evaluation of the temporal and spatial deviation from the reference value could help us mark out the areas in hazard due to the loss of biomass quantity. Deviations can be projected as the reactions of the vegetation to the potentially decreasing precipitation. Changes affecting large areas (forestry) were examined in the study period based on the CLC2000 and CLC2006 maps at a scale of 1:100 000. The spatial boundaries of surface-cover changes were taken into consideration when clarifying the results. The overlapping area of the changes in land cover (wood cutting, construction) and the low vegetation index result was low.

Characterising the status of vegetation for the period of 2000-2011

The biomass-potential can be best described by the 4-month period from May to August. The drought hazard is the most substantial in July and August, when the temperature is high. As for the average conditions (originally with low values), the mixed forests of Illancs region and a large part of the coniferous forests of the area south of Kecskemét are the most endangered areas. The minimum values map shows that the central and south-western part of Illancs, the northern part of Pilis-Alpári-homokhát (Pilis-Alpári Sandland) and that of the Kiskunsági-homokhát (Kiskunsági Sandland), the south and south-western part of Dorozsma-Majsai-homokhát (Dorozsma-Majsai Sandland) are the least drought-affected areas.

Signs of EVI reduction could be observed for 2001-2004 and for 2006-2009, but the period of 2006-2011 showed this more for mixed and pine forests! NDVI decrease for 2005-2009 could be stated for the pine. In spite of the impacts of 2001-2004 and 2010, the EVI and NDVI average values for coniferous forests have not shown a positive trend for 12 years. The favourable, more humid condition typical from March 2004 has shown by higher values at NDVI, but it dropped for the year of 2007. Even though 2010 broke the record in precipitation, it did not show remarkable results in vegetation. The water shortage characterising the area indicates that a dry year following years with more precipitation immediately reduces the biomass. By the increasing drought hazard, more situations that are similar to that of 2000, 2007 and 2011 could develop! The EVI, NDVI data for March-April for the deciduous trees can verify the analysis predicting that trees go green earlier due to warming up (the theory of longer vegetation period was not confirmed on the basis of the September data) (Fig. 6.27 on page 303).

The variability indicates the quick response of vegetation to the changes of the environmental conditions. Differences between the years could level off in a decade and could seem stable for a long term, however for a short time the period is extremely vulnerable, disposed to drought. This is clearly visible for the years of 2000, 2003, 2007, 2011 and for the months of April, July and August.

The study of deviation from the average in space and in time

The years of 2000, 2002, 2003, 2007, 2009 and 2011 (that is half of the years) showed negative deviations. The negative deviations of 2000 and 2003 for all forests are outstanding. The period of 2004-2006, three good years can be highlighted as a positive anomaly. Negative deviations characterise August-October, positive differences are more typical for April-May when studying the 16-day composites of EVI (Fig. 6.28 on page 304).

The highest negative and positive deviations occur in July-August and in spring. Both the greatest positive and negative deviation could be seen in a short period between July 2006 and July 2007! The areas showed negative values for the studied period, in spite of the averaging can be regarded as endangered from the viewpoint of climate change.

There is a substantial and an even higher negative EVI deviation altogether on 17,944 hectares, which is 14.1% of the studied area. By considering the different effects, the average values were calculated for the entire period for each cell, but cutting-down the woods, constructions in areas could have affected the appearance of negative values. There is a 20% overlap between the map of negative EVI deviation showing a rate of 14% and the map of permanent (construction) and temporary (forestry) loss of forests of anthropogenic origin. If the previous result is amended by this value, 11.3% is mostly endangered. 18% of Illancs forests are in danger, whereas there is a better situation in Bugaci-homokhát (Bugac Sand-land) (Fig. 6.29 on page 305).

The NDVI showed that 14% of deciduous forests are endangered. A more important negative deviation can be experienced on 5% of the areas. Deciduous trees are less imposed to danger as per NDVI than as per EVI. The mixed forests are in the worst situation, where 30% of the areas are noted for negative deviation. Positive anomaly characterises 20% of the areas. As for the total areas of deciduous forest, they show the greatest differences, but if separate area units are assessed, the mixed forests show the highest differences.

Summary

Chapter 3.3 and the above presented data confirm that satellite image datasets processed by geoinformatics could serve as good tools for the prediction of drought. The availability of higher resolution data could play an important role in the enhancement of results. The application of other drought indices, suitable to monitor the temporal development of drought within a year can also help the establishment of an early warning system based on vegetation condition.

The displayed results confirm that the study area is substantially affected by drought. Since climate simulations project the increasing occurrence of drought and an increase in extremities for the future, planning, management and relevant water management actions will be of vital importance not only for agriculture, but also for forestry and environmental protection.