A coupled climate-hydrological model for Meuse palaeodischarge modelling: set-up and calibration

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Philip J. Ward

Department of Palaeoclimatology and Geomorphology
Faculty of Earth and Life Sciences
Vrije Universiteit Amsterdam

Foreword

In this report the set-up and calibration of a coupled climate-hydrological model is described. The model will be used to simulate the response of the discharge of the Meuse River to millennial-scale changes in climate and land use. This research is being carried out in the framework of the Dutch National Research Programme *Climate changes Spatial Planning* (www.klimaatvoorruimte.nl), project CS9, entitled *Modelling and reconstructing precipitation and flood frequency in the Meuse catchment during the late Holocene*. The project is being carried out at the Department of Palaeoclimatology and Geomorphology of the Vrije Universiteit Amsterdam, in co-operation with the Institute for Environmental Studies (IVM) of the Vrije Universiteit Amsterdam, and TNO-NITG.

The data used in this project have many different sources. Discharge data have been obtained from Rijkswaterstaat RIZA (Netherlands), Direction Générale des Voies Hydrauliques Region Wallonne (Belgium), DIREN Lorraine Bassin Rhin-Meuse (France), and Roer and Overmaas Water Board (Netherlands). Climate data have been obtained from the Climate Research Unit (U.K.), and the European Climate Assessment & Dataset (ECA&D). Land use data have been provided by the European Environment Agency (Denmark). Soil data were obtained from the United Nations Food and Agriculture Organization (Italy), and a digital elevation model was obtained from the U.S. Geological Society (U.S.A.). All of these sources are gratefully acknowledged. The coupled climate-hydrological model described in this report is composed of two elements: the climate model ECBilt-CLIO-VECODE and the hydrological model STREAM.

I would like to thank Hans Renssen, Jeroen Aerts, Ronald van Balen, Jef Vandenberghe, Marcel de Wit, Frans Bunnik, Hans de Moel, Jos de Moor, and Laurens Bouwer for their valuable contributions.

Summary

In recent years the frequency and magnitude of high-flow events on the Meuse has been relatively great, and river flooding and flood risk mitigation have become major research themes. To date the vast majority of the research carried out on Meuse discharge has concentrated on records of the last century and simulations of the coming century. However, it is difficult to delineate changes caused by human activities and natural fluctuations on these timescales. Hence, we have set-up a coupled climate-hydrological model capable of simulating changes in daily discharge patterns on millennial timescales. This approach offers a new tool to assess the effects of anthropogenic changes in climate and land use on discharge, compared to the effects of natural climatic variations.

In this report we discuss the considerations used in the selection of the climate and hydrological models. We also describe in detail the set-up and calibration of the model. The hydrological model is calibrated for the relatively wet and warm period 1961-2000, and validated for the relatively dry and cool period 1921-1960. Modelled data are compared with independent observed data on discharge for the main Meuse and its tributaries, as well as data on evapotranspiration and snow-cover. In general, both the calibration and validation show good agreement between modelled and observed annual, monthly and daily discharge characteristics, except for the simulation of monthly and daily discharges in small tributaries. Hence, the model can be used to examine the effects of moderate climatic change on Meuse discharge. Model accuracy is not affected by the relative percentage cover of different land use types between subcatchments, and is therefore also suitable for investigating the effects of land use change. In order to facilitate this, we have created a GIS database of changes in land use and soil water holding capacity for the periods 4000-3000BP and 1000-2000 AD. The datasets form a useful resource for other members of the research community interested in the palaeohydrology and palaeoecology of the Meuse basin.

As the modelling approach is capable of simulating the effects of changes in climate and land use on Meuse discharge, and can be run efficiently for millennial-scale time-periods, it is well suited to the simulation of Meuse palaeodischarge over the late Holocene period, which will form the next phase of this research project.

Contents

1 Introduction

In recent years the frequency and magnitude of high-flow events in the Meuse basin (northwest Europe) (Fig. 1) has been relatively great compared to the rest of the 20th Century. As a result, river flooding and flood mitigation have become major research themes, and numerous studies have been carried out to examine the hydrological and climatological changes in the basin.

1.1 Previous studies of climate and discharge changes in the Meuse basin

The Royal Dutch Meteorological Institute (KNMI, 1994) states that the average temperature in the Netherlands was 0.7°C higher in the last 20 years of the 20th Century, compared to the first 20 years. Tu et al. (2004) carried out a statistical examination of the aerially averaged precipitation records of seven Belgian rainfall gauging stations for the period 1911-2002, and whilst they found no significant linear trend over the full period, a change-point analysis shows that the annual precipitation total has increased since 1980. More striking results are discernible for extreme precipitation events. For the period 1911-2002, Tu et al. (2005) found that winter extreme precipitation depths for periods of *k* days (*k*-day precipitation depths) over the Meuse basin (for *k* = 1, 3, 5, 7, 10 and 15) show an increasing trend with high probabilities (>90%). In an examination of observed precipitation data from Belgium for the period 1951-1995, Gellens (2000) also noted an upward trend in extreme winter *k*-day precipitation depths. Summer extreme *k*-day depths (for *k* = 5, 7, 10 and 15) show an abrupt increase at around 1980. From 1980 onwards the number of days (annual and winter half-year) with precipitation in excess of 10mm, along with the associated precipitation depths, increased significantly (Tu et al., 2005). Increased mean precipitation and intense rainfall events in the latter part of the 20th Century are related to fluctuations of large-scale atmospheric circulation (e.g. Bouwer et al., 2006) and a strengthened North Atlantic Oscillation (Tu, 2006). The results of simulations carried out with different General Circulation Models (GCMs) and Regional Circulation Models (RCMs) for the 21st Century tend to suggest that average temperatures and winter precipitation depths in the Meuse basin will increase, whilst summer precipitation will decrease (De Wit et al., 2007; Kwadijk & Rotmans, 1995; Pfister et al., 2004; Van Deursen, 2000a).

To relate these changes in climate to possible changes in the discharge regime of the Meuse, numerous studies have examined observed records of Meuse discharge over the 20th Century. Observed discharge time-series for Borgharen show no significant trends or change-points for annual or monthly average discharge over that period (De Wit et al., 2001; Tu et al., 2004; WL, 1994a). However, over the last century, increasing trends (albeit statistically insignificant) are detected for annual maximum daily discharge (Tu et al., 2005) and winter maximum daily discharge (De Wit et al., 2001; Tu et al., 2005; WL, 1994a). Furthermore, change-point analyses suggest a statistically significant increase in annual and winter maximum daily discharge at about 1984 (Pfister et al., 2000; Tu et al., 2004, 2005). Similar results have been found for a number of

Meuse tributaries (Tu et al., 2005). Low-flows on the Meuse (e.g. annual minimum daily discharge and annual minimum 10-day discharge) show a significant downward trend over the period 1911-2002 (De Wit et al., 2001; Tu et al., 2004). The results of these studies therefore suggest that the mean monthly and annual discharges have remained relatively stable over the 20th Century, whilst the variability and extremes have increased.

Research into possible future changes in Meuse discharge, based on hydrological modelling, shows a continuation of this pattern. De Wit et al. (2001) suggest that mean annual discharge will remain relatively stable over the 21st Century, although Booij (2005) suggests a slight decrease. The seasonal distribution of discharge may change so that winter discharge becomes greater and summer discharge lower (De Wit et al., 2001, 2007). These and other studies, however, suggest that the anticipated climate change of the 21st Century will lead to an increase in flood frequency, especially in the winter half-year (Booij, 2005; Bultot et al., 1988; Gellens & Roulin, 1998; Middelkoop & Parmet, 1998; Van Deursen, 2000a). Fewer hydrological modelling studies have specifically examined the effects of future climate change on low-flows, but those which have been carried out suggest an increase in the frequency and magnitude of low-flow events (Booij, 2005; Bultot et al., 1998; De Wit et al., 2001, 2007).

Discharge, however, depends not only on climate but also on land use. Little research has been carried out to examine the effects of land use change in the Meuse basin because the forested area has remained relatively stable at the basin-scale over the course of the last century (DGRNE, 2000; Knol et al., 2004; Tu, 2006). Ashagrie et al. (2006) and Tu (2006) found no evidence to suggest that land use changes in the 20th Century have had a significant effect on Meuse discharge.

1.2 Rationale behind the current research

The aforementioned studies have examined either the discharge changes of the last century, or used hydrological models calibrated against observations for that period to simulate changes in the 21st Century. However, when studying long-term discharge changes (centennial to millennial) this is problematic. Firstly, accurate daily measurements of Meuse discharge have only been made since 1911, which is too short to evaluate long-term climatic changes (Alcamo et al., 2000; Meybeck, 2003). Secondly, at a basin-scale the forested area has been relatively stable over the last century, so it is difficult to assess the effects of large-scale changes in land use.

Studies of palaeodischarge provide a means to address this lack of long-term observed data by providing a dataset for the validation of model response on millennial timescales. This enables us to simulate the response of discharge to changes in climate over longer timescales, to greater amplitudes of climatic change, and to more extensive changes in land use. Furthermore, the use of millennial timescales allows for a more realistic investigation of extreme events with low return frequencies. However, palaeodischarge modelling is still in its infancy. For example, Bogaart et al. (2003) used a process-based numerical model to simulate long-term (> 100 year) changes in mean Meuse discharge during the last Glacial-Interglacial transition. Coe and Harrison (2002) used runoff derived directly from a GCM, in combination with a river routing algorithm, to simulate lake level changes in northern Africa at ca. 6 ka BP. At the basin-scale, however, the use of runoff data derived directly from GCMs has a disadvantage since runoff is less well resolved than climate. To address this problem, Aerts et al. (2006) and Ward et al. (2007) coupled a climate model and a hydrological model to simulate the Holocene discharge of nineteen rivers around the globe on a monthly time-step; the modelled discharges correspond well with multi-proxy records of palaeodischarge.

Hence, in this research project a coupled climate-hydrological model was set-up to simulate the daily discharge of the Meuse using the modelling approach of Aerts et al. (2006) and Ward et al. (2007), but with a daily time-step and higher spatial resolution. Meuse discharge will be simulated for two time-slices: 4000-3000 BP and 1000-2000 AD. The period 4000-3000 AD is selected as a natural reference period since the natural climatic forcings were broadly similar to those of today, whilst human influence on land use was minimal (Bunnik, 1995; Gotjé et al., 1990; RWS Limburg/IWACO, 2000). The period 1000-2000 AD was heavily influenced by human activities, namely changes in land use throughout the period, and greenhouse gas emissions since the industrial revolution. By comparing the changes in discharge characteristics between these two situations the effects of anthropogenic changes in climate and land use can be examined.

1.3 Study area

The Meuse is a predominantly rain-fed river with a total length of ca. 875 km from its source in France to its outlet in the Netherlands (Fig. 1). The catchment extends over parts of Belgium, France, Germany, Luxembourg, and the Netherlands, and has an area of ca. 33,000km^2. For this project the model was set up upstream from Cuijk (Fig. 1); downstream from this point it becomes increasingly difficult to separate the Meuse and Rhine systems.

From a hydrological perspective the Meuse can be split into three main sections: (a) the upper reaches (Lotharingian Meuse); (b) the central reaches (Ardennes Meuse); and (c) the lower reaches (Dutch Meuse). The Lotharingian Meuse extends from the source at Pouilly-en-Bassigny (France) to the confluence with the Chiers (France) (ca. 25 km downstream from Stenay), and is characterised by a lengthy and narrow catchment of low gradient through mainly sedimentary Mesozoic rocks; along this section there are few major tributaries. The Ardennes Meuse extends from the confluence of the Meuse and the Chiers (France) to the Dutch border near Eijsden (ca. 10 km south of Maastricht); along this reach the major tributaries Viroin, Semois, Lesse, Sambre and Ourthe join the Meuse. The Ardennes Meuse transects mainly Palaeozoic rocks of the Ardennes Massif and the river is narrow and steep. The permeability of the soils in much of this area is low and this, in combination with the steep gradient of the Meuse and its tributaries, encourages a quick response to precipitation. The Dutch Meuse extends from Eijsden to the confluence with the Rhine in the Hollandsche Diep. The gradient in the upper part of this section is still relatively steep, but the gradient decreases downstream from Maastricht, where the landscape mainly consists of unconsolidated Cenozoic sediments. The main tributaries in this section are the Roer, Niers and Dieze (Berger, 1992).

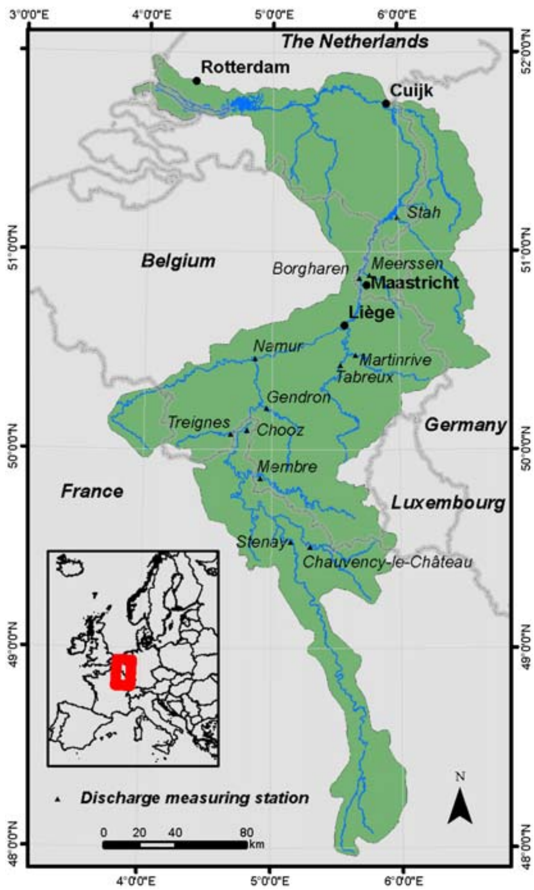

Fig. 1: Map showing the location of the Meuse basin and discharge measuring stations used in this study (Data source: RWS Limburg/IWACO, 2000). The inset shows the location of the Meuse basin in Europe.

The mean annual precipitation over the Meuse basin is ca. 950 mm, and is reasonably evenly distributed throughout the year. The spatial distribution of precipitation is to a large extent a reflection of elevation and distance from the coast. Mean air temperatures show a marked seasonal variation, and annual potential evapotranspiration is much greater in the summer half-year than in the winter half-year (accounting for 76% and 24% of the total respectively) (Ashagrie et al., 2006). The mean annual discharge of the Meuse and its associated canals at the border of Belgium and the Netherlands is ca. 276 m^3s^{-1}; summer and winter half-year mean discharges are 146 m^3s^{-1} and 406 m^3s^{-1} respectively (Ashagrie et al., 2006).

2 Model selection and design

2.1 Overview of the modelling approach

The coupled climate-hydrological model described in this report is composed of two elements: the climate model ECBilt-CLIO-VECODE (Brovkin et al., 2002; Goosse and Fichefet., 1999; Opsteegh et al., 1998) and the hydrological model STREAM (Aerts et al., 1999).

The model was calibrated and validated against observed discharge records for the 20th Century, and will be run on a daily time-step for the periods 4000-3000 BP and 1000-2000 AD, as well as for a number of deforestation scenarios. For each time-slice the climate model will be run in ensemble mode, with four ensemble members (see Section 3.1). The simulation results will be compared to independent proxy records of Meuse discharge to assess the validity of the model in palaeodischarge retrodiction, and assessments made of the long-term effects of climatic and land use change on Meuse discharge over the late Holocene. Finally, the model will be used to simulate changes in discharge using future climate and land use change scenarios, thereby facilitating a comparison of discharge sensitivity to natural long-term climatic variations and to changes caused by anthropogenic activities. A schematic overview of the research approach is given in Fig. 2.

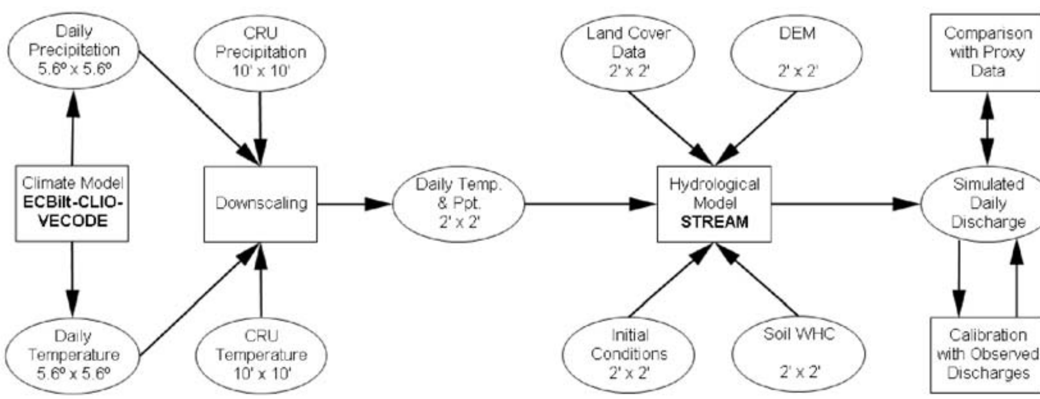

Fig. 2: Flowchart of the research approach and methodology to be used in this study.

2.2 Climate model selection and specifications

Traditionally the results of GCMs have been widely employed in hydrological modelling: examples from northwest Europe include the studies of De Wit et al. (2001), Gellens and Roulin (1998), Middelkoop et al. (2001), Pilling and Jones (1999), and Sefton and Boorman (1997). However, since the resolution of these models is typically at least 2.5° x 2.5° (Bouwer et al., 2004), they are rather coarse for direct use in regional hydrological studies (Arnell et al., 1996; Bouwer et al., 2004; Kleinn et al., 2005; Wood et al., 2002, 2004). Hence, numerous recent studies have used the results of higher resolution regional circulation models (RCMs) to force hydrological models (e.g. De Wit et al., 2007; Kleinn et al., 2005; Shabalova et al., 2003).

However, the high computational cost of running GCMs and RCMs makes their use in hydrological studies on millennial timescales unfeasible. Therefore, in this study, ECBilt-CLIO-VECODE (Brovkin et al., 2002; Goosse and Fichefet, 1999; Opsteegh et al., 1998) is used, an Earth System Model of Intermediate Complexity (EMIC). EMICs include the majority of processes described in GCMs, but in a more reduced, or parameterised, form. This simpler nature allows their application to long-term climate simulations over periods of several thousands of years (Claussen et al., 2002).

2.2.1 ECBilt-CLIO-VECODE

ECBilt-CLIO-VECODE is a three-dimensional coupled climate model consisting of three components describing the atmosphere, ocean and vegetation. The atmospheric component (ECBilt2) is a T21, 3-level quasi-geostrophic model (Opsteegh et al., 1998). The ocean-sea-ice component (CLIO) consists of an ocean general circulation model coupled to a thermodynamic-dynamic sea-ice model (Goosse and Fichefet, 1999). The final component is VECODE, a dynamic vegetation model which simulates the dynamics of two main terrestrial plant types (forest and grasses) as well as bare soil, in response to climatic fluctuations (Brovkin et al., 2002).

2.3 Hydrological modelling

There exists a plethora of models to simulate the hydrological cycle at a range of temporal and spatial resolutions. The concepts employed in hydrological models reflect the application for which the model will be used, and since the range of applications is large, so too is the range of modelling concepts. However, in their simplest form all hydrological models are governed by the water balance equation:

$$Q = P - ET \pm \Delta S \qquad (1)$$

where Q is discharge, P is precipitation, ET is evapotranspiration and ΔS is the change in storage in the system.

Hydrological models can be classified according to many different criteria, including model application (e.g. within hydrology as exploratory or research tools for the understanding of hydrological processes, or outside hydrology in planning, design and prediction), model structure (from simple lumped black-box models to complex distributed physical and mathematical models), spatial disaggregation (lumped or distributed parameters), computational demand, temporal scale (e.g. hourly, daily, monthly, annual), and spatial scale (e.g. point, sub-basin, basin, regional, global) (Booij, 2005; Klemeš, 1986; Leavesley, 1994). Hereunder follows a brief description of the main modelling approaches based on model structure, after Leavesley (1994).

Empirical models: Such models are based on statistical relationships between components of the water balance equation, and convert input to output via a transform function statistically derived from empirical research. Hence they give no consideration to the physical laws governing the processes involved in the hydrological cycle (Leavesley, 1994). Whilst these models may be of use in the rapid assessment of contemporary rainfall-runoff relationships their application to long-term studies, or studies involving changes in major parameters (i.e. climate and land use) is therefore questionable (Kirchner, 2006).

Water balance models: Water balance models are based on the concepts and works of Thornthwaite (1948) and Thornthwaite and Mather (1955), and use a series of storage compartments and flows to account for the movement of water throughout a basin. Most models of this type consider total discharge to be a function of a delayed runoff component (baseflow) plus a direct runoff component. They are based on simple representations of hydrological processes, and given their quasi-physical basis these models can be used to simulate average discharge for a range of basin conditions, since they allow for the parameterisation of changes in variables such as precipitation, evapotranspiration, land use, and soil type. Parameter estimation is often difficult given the lack of empirical observations on such variables as groundwater and soil water storage. However, only a limited number of parameters are considered in these models, which is advantageous over process-based distributed-parameter models (see below). Whilst the equations used can offer a robust tool for assessing long-term discharge patterns, they are too primitive to be able to model hydrological responses on timescales smaller than a day (Leavesley, 1994).

Conceptual lumped-parameter models: Similar to water balance models, conceptual lumped-parameter models attempt to account for the flow of water through a basin from the moment it enters as precipitation until the moment it is discharged at the outlet. They are also based on approximations and simplifications of physical laws, although flow paths and residence times in the storage compartments are considered in far greater detail. Hence, they tend to be applied with much shorter time-steps, for example minutes, hours, or a day (Leavesley, 1994). Process parameterisation usually takes place at the basin-scale, and heterogeneous factors such as soil type and land use tend to be lumped at this scale. Therefore it is difficult to use these models for spatial planning, for which the alteration of local conditions needs to be represented.

Process-based distributed-parameter models: These models have a firm grounding in physical processes and in the processes involved in the hydrological cycle at the individual basin-scale. The process equations have the ability of forecasting the spatial pattern of hydrological conditions as well as basin-wide storage and flows (Beven, 1985). Unlike conceptual lumped-parameter models, spatial distribution is applied using a raster approach whereby the process equations are carried out for each individual grid cell. However, as the physical reality of a model increases, so too does the number of parameters which must be estimated. Since empirical data on which to base such parameter estimations is usually scarce, and sometimes non-existent, over-parameterisation introduces many uncertainties into these models (Jakeman and Hornberger, 1993; Krajewski et al., 1991; Loague and Freeze, 1985). When the purpose of a model is to simulate the detailed processes of the hydrological cycle, such as in models for water quality management (Zheng and Keller, 2006), this high number of parameters may be necessary. However, unnecessary over-parameterisation often leads to little (if any) improvement in model performance (Beven, 1989; Booij, 2005; Jakeman and Hornberger, 1993; Loague and Freeze, 1985).

2.3.1 Desired characteristics of Meuse palaeodischarge model

Given the issues discussed above, the following points were considered during the selection of the hydrological model to be used in this project:

- the model should be spatially distributed in order to simulate changes in land use;
- the model must be able to deal with altered climatic conditions, hence a purely empirical model is inappropriate. Furthermore, our goal is to better understand the main processes which govern the response of discharge to climate change;
- the model must be capable of modelling processes on a daily basis in order to simulate individual high-flow and low-flow events;
- the spatial disaggregation and resolution must be high enough to allow effective daily discharge modelling at the basin-scale and for the main tributaries;
- the model should contain as few calibration parameters as possible, whilst still retaining a physical grounding, given that parameter uncertainty in the palaeodischarge runs adds to overall model uncertainty;
- the model must be efficient to run in terms of computational time, given that a single 1000-year run involves 360,000 iterations.

Based on these considerations, the most appropriate model structure for this project is one combining elements of the water balance models and the process-based distributed-parameter models.

2.3.2 STREAM

The STREAM model (Aerts et al., 1999) fits these criteria well, and was therefore selected for use in this project. STREAM is a grid-based spatially distributed water balance model that describes the hydrological cycle of a drainage basin as a series of storage compartments and flows. It is based on the RHINEFLOW model of Kwadijk (1991, 1993). STREAM calculates the water balance per grid cell using the Thornthwaite equations for potential evapotranspiration (Thornthwaite, 1948) and the Thornthwaite & Mather equations for actual evapotranspiration (Thornthwaite and Mather, 1957). The main flows and storage compartments used to calculate water availability per grid cell are shown in Fig. 3. The direction of water flow between grid cells is based on the steepest descent for the eight surrounding grid cells on a digital elevation model.

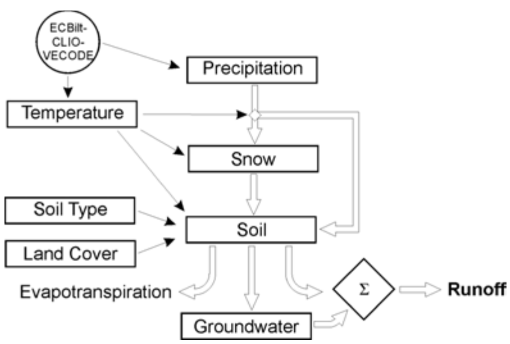

Fig. 3: Flowchart showing the main storage compartments and flows of the STREAM palaeodischarge model (Ward et al., 2007).

STREAM (or its predecessor RHINEFLOW) has been successfully applied to numerous basins of varying sizes in different parts of the world for studies of the 20th and 21st Centuries, e.g. the Rhine in Europe (Van Deursen and Kwadijk, 1994a); the Ganges-Brahmaputra and Krishna in India (Bouwer et al., 2006; Van Deursen and Kwadijk, 1994b); the Yangtze in China (Van Deursen and Kwadijk, 1994b); the Perfume River in Vietnam (Aerts and Bouwer, 2002a). Furthermore, Aerts et al. (2006) and Ward et al. (2007) applied the model to numerous large basins around the globe to simulate monthly discharge over the course of the Holocene, and found the results to be in good agreement with proxy evidence of changes in palaeodischarge.

2.3.3 Selection of resolution for the STREAM model

The STREAM model was set up on a daily time-step and at a spatial resolution of 2’ x 2’ (ca. 2.4 km x 3.7 km). The considerations for the choice of these time and spatial resolutions are discussed in this section.

As this research project aims to examine changes in both mean discharge and extreme events over the Holocene, a daily time-step is appropriate (Booij, 2002a). In terms of modelling the response of the Meuse to long-term climatic and land use changes, a higher time resolution (e.g. hourly) is unnecessary since we are interested in the change in the number of extreme events and average discharge, rather than the detailed response of the hydrograph.

A version of RHINEFLOW (predecessor of STREAM) has previously been set up for the Meuse (MEUSEFLOW; Van Deursen, 1999, 2000a), and run at a scale of 1 km x 1 km. It was initially intended to run the Meuse palaeodischarge model at the same spatial resolution, for comparative purposes. However, the majority of the datasets used as input in the palaeodischarge model have a geographical projection (i.e. are in degrees latitude and longitude). Hence, to improve the efficiency of data conversion, and to reduce the introduction of errors due to reprojection, the palaeodischarge model was set up in geographical units. The choice of spatial scale should allow a model to capture the dominant processes, yet not be so complex that excessive computational time is demanded (Booij, 2003). Based on this notion, Booij (2002a) calculated the "appropriate" model scale for assessing the impact of climate change on flooding in the Meuse by firstly establishing appropriate spatial scales of key variables, assuming these to be equal to a fraction of the spatial correlation length of those variables. These appropriate variable scales were then weighted, and the results multiplied to establish an appropriate model scale of 10 km x 10 km.

Given the choice of a daily time-step it was important to select a pragmatic spatial resolution in terms of computational time and disk storage space. Therefore, a pilot model was set-up in order to assess the feasibility of three spatial resolutions at least as high as the 10 km x 10 km proposed by Booij (2002a), namely 30" x 30", 1' x 1' and 2' x 2'. For each resolution the total disk space and time which would be needed to carry out all of the simulations required in this project was estimated (Table 1). These times exclude the time taken to run the climate model itself, since this is not affected by the resolution chosen for the hydrological model. Based on this pilot study, pragmatic considerations concerning time and disk space dictated that the resolution of 2' x 2' be chosen.

Table 1: Disk space and calculation time requirements to run the model in ensemble mode (4 ensemble members) with a daily time-step for the periods 4000-3000 BP and 1000-2000 AD, and with multiple land use scenarios.

Resolution	Disk space	Calculation time
30" x 30"	ca. 4608 GB	ca. 20,640 hours (860 days)
1' x 1'	ca. 1152 GB	ca. 4240 hours (177 days)
2' x 2'	ca. 288 GB	ca. 1680 hours (70 days)

3 Model input data

To set up the model a GIS database of input maps was created, using the IDRISI Kilimanjaro software. The various input data files are described in this section.

3.1 Climate data

The input climate data (daily temperature and precipitation) were derived from the ECBilt-CLIO-VECODE model, and subsequently downscaled to the resolution of the STREAM model. The climate data used in this study were derived from a transient run, forced by annually varying orbital parameters and atmospheric greenhouse gas concentrations (CO_2 and CH_4) following Renssen et al. (2005), and atmospheric volcanic aerosol content and fluctuations in solar activity following Goosse et al. (2005). The climate model was run in ensemble mode, with four ensemble members. Each ensemble member represents a single model run. The ensemble members are forced using the same climatic parameters, but with slightly different initial climatic conditions to account for the chaotic behaviour of the atmospheric system. Hence, the difference between the ensemble members gives an idea of the natural variability within the system, whereas an ensemble mean can be used to evaluate long-term trends.

The use of a climate model to simulate precipitation on a daily time-step has an important limitation; climate models underestimate precipitation variability on a daily time-step, since they have a low resolution and simplified physics. Weather simulation models are better at simulating the daily variability of regional and local precipitation. However, they are much more complicated and time consuming to run, and are not designed to simulate weather patterns under conditions of greatly changed climates. Hence, their use for climate change studies is neither feasible nor desirable. Rainfall generators can be used to force precipitation in hydrological studies, but are also based on mathematical representations of present day weather frequency distributions. Altering the frequency distribution probability functions for future or past climates assumes an a priori knowledge of past or future changes in precipitation, whilst these changes are in fact one of the fundamental aspects being researched. A climate model uses forcing parameters deduced from empirical studies to simulate changes in the past based on those parameters. Whilst it is appreciated that daily precipitation depths simulated by climate models tend to underestimate actual variability, their ability to simulate long-term changes and their efficiency in terms of run time make them the most appropriate method for this project.

3.1.1 Downscaling

The climate data derived from ECBilt-CLIO-VECODE have a spatial resolution of ca. 5.6° x 5.6°, and were therefore downscaled to the resolution of the STREAM model. The first step was a spatial downscaling procedure (Bouwer et al., 2004), whereby the values from the ca. 5.6° x 5.6° climate model grid were simply resampled onto a 2' x 2' grid.

The resulting spatially downscaled climate maps have a spatial resolution of 2' x 2', but only reflect spatial variability at the scale of the climate model (ca. 5.6° x 5.6°). Climate model data at this scale are too coarse to be used directly in regional and basin-scale hydrological studies (Arnell et al., 1996; Bouwer et al., 2004; Kleinn et al., 2005; Wood et al., 2002, 2004). Therefore a second downscaling step was required to introduce a more realistic and greater degree of spatial variability. Bouwer et al. (2004) identify two main approaches: statistical methods which transform the data in such a way as to match the main statistical properties of modelled and observed climate data sets (e.g. Bouwer et al., 2004; Wilby and Wigley, 1997; Wilby et al., 1998; Wood et al., 2002); and dynamical approaches which use finer resolution regional circulation models (RCMs) nested within coarser GCMs (e.g. Cocke and LaRow, 2000; Kim et al., 2000; Murphy, 2000; Wood et al., 2002; Yarnal et al., 2000). The results of these two approaches have been found to have similar levels of skill (Wilby et al., 2000; Wood et al., 2004) but dynamical methods are computationally far more demanding (Bouwer et al., 2004). The number of daily iterations involved in this project renders the latter approach prohibitive.

Statistical downscaling involves the use of correction factors (for temperature additive and for precipitation multiplicative) which are applied to the low resolution model data so as to preserve the statistical properties of a higher resolution observed (baseline) dataset. In this study the spatially explicit correction method based on monthly averages as described by Bouwer et al. (2004) was used. For the observed baseline datasets of precipitation and temperature the CRU TS 1.2 dataset (Mitchell and Jones, 2005) was used. This dataset has a spatial resolution of 10' x 10', which is similar to the appropriate scale of 20 km for precipitation data calculated by Booij (2002b, 2003).

The spatially downscaled ECBilt-CLIO-VECODE data were spatially redistributed using the following formulae:

$$p'_{ECB(d,i)} = p_{ECB(d,i)} \times \left(\frac{\overline{p}_{CRU(m,i)}}{\overline{p}_{ECB(m,i)}} \right) \qquad (2)$$

where $p'_{ECB(d,i)}$ is the spatially redistributed ECBilt precipitation for a particular day, d, and cell, i, $p_{ECB(d,i)}$ is the raw ECBilt precipitation for a particular day, d, and cell, i, $\overline{p}_{CRU(m,i)}$ is the observed (CRU) average monthly precipitation for a particular month, m, and cell, i, and $\overline{p}_{ECB(m,i)}$ is the raw ECBilt mean monthly precipitation for a particular month, m, and cell, i.

$$t'_{ECB(d,i)} = t_{ECB(d,i)} + (\bar{t}_{CRU(m,i)} - \bar{t}_{ECB(m,i)}) \qquad (3)$$

where $t'_{ECB(d,i)}$ is the spatially redistributed ECBilt temperature for a particular day, d, and cell, i, $t_{ECB(d,i)}$ is the raw ECBilt temperature for a particular day, d, and cell, i, $\bar{t}_{CRU(m,i)}$ is the observed average monthly temperature for a particular month, d, and cell, i, and $\bar{t}_{ECB(m,i)}$ is the raw ECBilt mean monthly temperature for a particular month, d, and cell, i.

3.1.2 Downscaled climate data

The downscaled mean monthly temperature and precipitation data for the period 1901-2000 show very good agreement with observed values (CRU) for all ensemble members (Fig. 4) (temperature, *r*>0.99; precipitation, *r*>0.96).

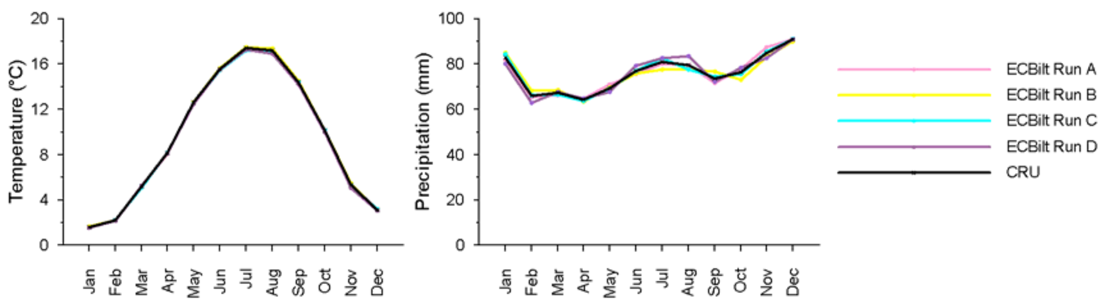

Fig. 4: Correlation between observed (CRU) and simulated (ECBilt) mean monthly basin temperature (left) and precipitation (right) for the period 1901-2000. The simulated data are shown separately for each ensemble member (A, B, C, D). In all cases the correlation between observed and simulated temperature and precipitation is statistically significant (temperature, r>0.99; precipitation, r>0.96).

The frequency distributions of monthly mean temperatures for the simulated and observed datasets also show statistically significant similarity according to both the Kolmogorov-Smirnov (KS) test (α=0.05), and the Mann-Whitney U test (MWU) (α=0.05). For mean monthly precipitation the observed and simulated frequency distributions show a similar measure of central tendency (MWU-test, α=0.05), yet the simulated dataset shows less variability than the observed dataset (especially in spring), due to the limitations mentioned in Section 3.1.

The downscaled daily temperature and precipitation data were validated against observed values for Maastricht, from the European Climate Assessment & Dataset (ECA&D) (http://eca.knmi.nl/)

(Klein Tank et al., 2002), for the period 1911-2000. It should be noted that even under a situation where modelled data were perfectly downscaled to fit the CRU data, discrepancies would still be expected when comparing the downscaled modelled data to observations at single weather stations. This is because the CRU data, on which the downscaling is based, have a spatial resolution of 10' x 10', corresponding to a raster cell area of ca. 221 km^2, whereas the weather station data refer to a single point. The frequency distributions of observed and simulated daily temperature show a statistically similar measure of central tendency (MWU-test, α=0.05). However, the simulation results show a significant underestimation of the number of days on which mean temperature is between ca -2 and +2 °C (Fig. 5). As this discrepancy could have an effect on the snowfall and snowmelt characteristics of the basin, those parameters in the model pertaining to snow characteristics have been adjusted accordingly (see Section 5.2). The downscaled climate model data slightly underestimate the occurrence of warm days between ca. 22° and 26°C, and overestimate the number of days with temperatures between ca. 16° and 22°C. The magnitudes of these over and underestimations are relatively small, and will act in a compensatory fashion in the estimation of evapotranspiration.

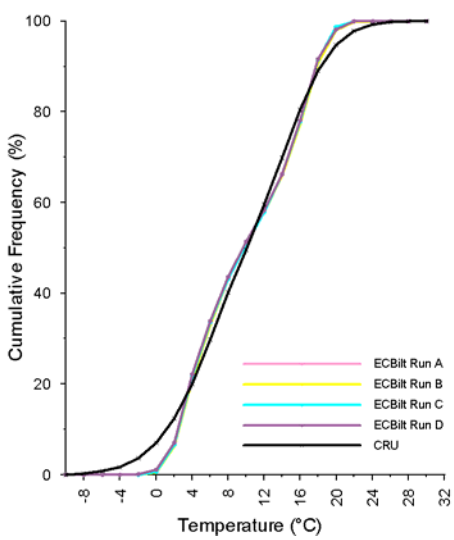

Fig. 5: *Cumulative frequency distributions of observed (CRU) and simulated (ECBilt) mean daily temperatures at Maastricht (1911-2000), showing reasonable overall agreement between the two datasets.*

The number of precipitation-free days at Maastricht (model, 27.5%; observed, 45.7%) is underestimated. This is a common problem in daily climate modelling, and is partly related to the simplified physics used to describe atmospheric processes, and partly related to the fact that the original model grid cell represents an area of ca. 5.6° x 5.6°, whilst the observed data refer to a specific point. Since the spatial downscaling method used here redistributes precipitation over the basin according to long-term average patterns, it means that convectional rainfall is poorly simulated, whereas frontal rainfall is much better resolved. In reality, the majority of floods on the Meuse are the result of frontal precipitation over large parts of the basin over extended periods of times, especially under conditions of antecedent soil saturation (Berger, 1992). No significant difference was found between the frequency distributions of observed and simulated daily precipitation for very wet days (10% wettest days) (Kruskal-Wallis (KW) test, χ^2 = 1.691, p=0.792).

3.1.3 Evapotranspiration

For each grid cell, STREAM calculates potential evapotranspiration (PE) based on the temperature-dependent equations of Thornthwaite (1948), and actual evapotranspiration (AE) based on the Thornthwaite and Mather equations (1957). Adaptations of these equations, as well as those used to calculate the water balance per grid cell, can be found in Appendix 1. Hence, the values of AE and PE are dependent on the calibration parameters used, and are therefore discussed in Section 5.2. This approach to AE and PE estimation was used in order to simulate the spatial variability of these variables throughout the basin. ECBilt-CLIO-VECODE also produces daily PE and AE data, but these values are not used here since no high resolution spatially distributed observed datasets of AE are available for downscaling these to the 2' x 2' resolution of STREAM.

3.2 Flow direction map

The flow direction map is used in STREAM to determine the direction of water flow between grid cells (Kwadijk, 1993; Van Deursen 1995; Van Deursen and Kwadijk, 1994a), based on the steepest descent for the eight surrounding grid cells on a digital elevation model (DEM). In this project the U.S. Geological Survey (USGS) GTOPO30 DEM (tile W020N90) was used (http://edc.usgs.gov/products/elevation/gtopo30/gtopo30.html).

Using the flow direction map, the drainage basin upstream from Cuijk in the Netherlands was calculated in IDRISI. The agreement between the two drainage networks is very good; at Borgharen the modelled area being just 1.7% smaller than the measured area. However, at the resolution used in this model, the Cuesta of Florenville (a ridge separating the Semois from the Chiers) is poorly represented due to the resolution of the original DEM (10' x 10'), and hence the Semois and Chiers tributaries are incorrectly delineated. Hence, the Chiers and Semois will only be used in further discharge analyses when their discharge totals are combined.

3.3 Crop factor maps

A crop factor map is used in STREAM for the calculation of PE. Different crop factors are applied to different land use types to account for their relative differences in PE. For this study a crop factor map was prepared for each century by first reconstructing a land use map for each century, and then reclassing these to crop factors. Given the relative scarcity of detailed historical land use data, a simplified set of land use categories was used, and these were reclassed to crop factors based on values in existing studies (Kwadijk, 1993; Aerts and Bouwer, 2002b): Urban (0.8), Forests (1.1), Agriculture and Grasslands (0.9), Wetlands (1.1), Water Bodies (1.5). The reconstructed land use maps for each century can be found in Appendix 2.

3.3.1 Recent land use

For 20th Century land use the CORINE Land Cover 2000 dataset (CLC2000, 250m version 8/2005 (V2)) (© EEA, Copenhagen, 2005) was used. This dataset is available at http://dataservice.eea.eu.int/dataservice/, and is based on the photo-interpretation of satellite images.

3.3.2 Land use 1000-2000 AD

For the Meuse basin, a wealth of information on land use change is available from an amalgam of sources (e.g. reviews of historical records, pollen analyses, archaeological studies), and was employed in the reconstruction of land-use over the last millennium.

11th Century: As part of a major study to assess the ecological functioning of the Meuse at the basin-scale, RWS Limburg/IWACO (2000) compiled a historical review of the catchment's ecological functioning (including land use change) over the last millennium. Based mainly on the works of Dornbusch (1997) regarding the extent of forest and important settlements in Europe around 1000 AD, and of Van Es et al. (1988) regarding the distribution of peat bogs and fenlands at that time, RWS Limburg/IWACO (2000) produced a map of Meuse basin land use at ca. 1000 AD (Fig. 6). The data in the map are supported by the results of independent studies from other sources (e.g. Blouet et al., 1993; Buisman and Van Engelen, 1995; Dierkens and Plumier-Torfs, 1999).

In this map, land use is grouped into eight categories (Fig. 6). For the present study the map was rasterised and reprojected, and the land uses were reclassed as follows:

- *Dunes:* do not occur in the part of the catchment upstream from Cuijk;
- *Peat bogs*: reclassed to wetlands;
- *Fenlands:* large areas of fenland would have been covered by fenland forest at this time (RWS Limburg/IWACO, 2000), and hence half of the cells of this class were reclassed to forests, whilst the other half were reclassed to wetlands. Given that forests and wetlands have the same crop factor (1.1) this approach is acceptable;
- *Primary deciduous forests*: reclassed to forests;
- *Floodplains*: this category refers to a mosaic land cover made up for the main part of grassland pastures and meadows, but with a patchwork of willow riverine forest. In the present study two-thirds of these cells were reclassed to agriculture and grasslands, with every third cell being reclassed to forests;
- *Tidal marshes and mudflats:* do not occur in the part of the catchment upstream from Cuijk;

- *Vineyards*: reclassed to agriculture and grasslands;
- *Cultivated land*: reclassed to agriculture and grasslands.

Since the land use map of 1000 AD does not contain any information on water bodies, this class is held constant throughout the entire study period, based on its present-day distribution.

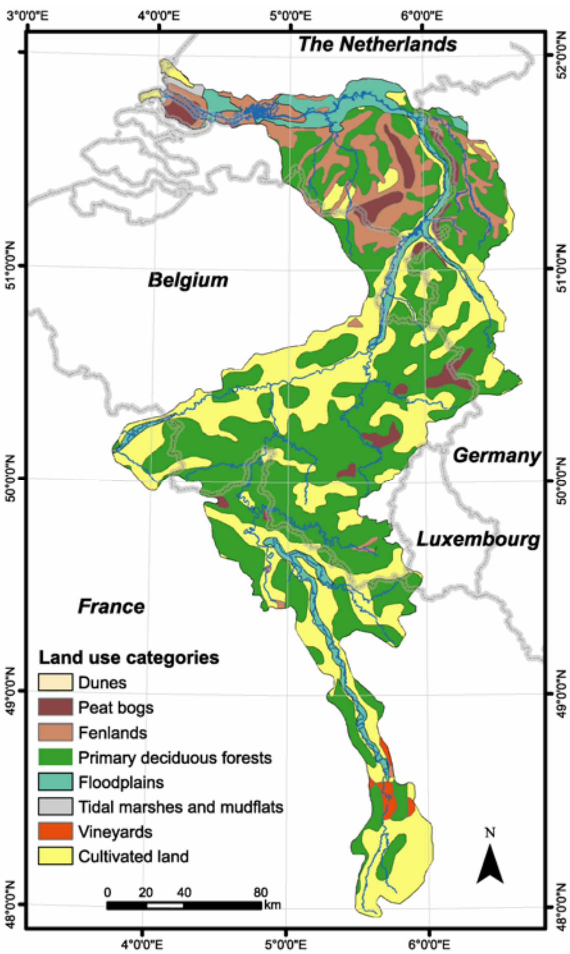

Fig. 6: Land use in the Meuse basin at ca. 1000 AD (RWS Limburg/IWACO, 2000).

19th Century: For the 19th Century, data on forest cover and urban area were based on historical census information for Belgium (WL, 1994), and on historical maps for the Netherlands (Knol et al., 2004). Data on wetlands and peat bogs were based on the works of Damblon (1992), Petit and Lambin (2002), and RWS Limburg/IWACO (2000). For the French subcatchments census data on forest cover was used, based on the work of Dutoo (1994). For the German and Luxembourg sections of the basin the relative proportions of each land use type were extrapolated based on the values for Dutch Limburg and the Chiers subcatchment respectively. The area under wetlands was estimated based on data in RWS Limburg/IWACO (2000).

12th-18th Century: For this period no quantitative assessments of land use are available and instead the changes were estimated per subcatchment based on qualitative reviews of historical land use change:

- *Water Bodies*: assumed to remain constant;
- *Urban*: urbanisation was not a factor of importance until ca. 1850 AD (RWS Limburg/IWACO, 2000);
- *Forests*: between the 11th and 18th Centuries forests were exploited for fuelwood, charcoal, and building materials, and removed to make way for pastures. Later in the period wood was used as a fuel in the foundries of industrial regions (Froment, 1975; Gotjé et al., 1990; Petit and Lambin, 2002; RWS Limburg/IWACO, 2000; WL, 1994b). Although this forest degradation was concentrated in different parts of the basin during different periods, overall the period 1000-1800 AD was typified by a basin-wide decrease in forest. In the present study a linear reduction in forested area was assumed between the distributions of the 11th and 19th Centuries, as was assumed to be the case for Europe in the period 1000-1700 AD by Brovkin et al. (1999);
- *Wetlands*: fenland destruction commenced in the 16th Century (RWS Limburg/IWACO, 2000), and by the 19th Century almost complete destruction had occurred. Hence, in the present study it was assumed that a linear reduction of area covered by fenlands occurred between these periods. The destruction of peat bog appears to have commenced in earnest from the end of the 16th Century (RWS Limburg/IWACO, 2000), with a peak around the 18th Century, and the current situation being reached by the late 19th Century. Hence, in the present study it was assumed that the situation remained stable until the 16th Century inclusive, and that thereafter areal reductions occurred as follows: 17th Century (25% reduction), 18th Century (50% reduction), 19th Century (100% reduction). An exception to this pattern was in the Peel region (Netherlands), where peat bogs were intact until far into the 19th Century (RWS Limburg/IWACO, 2000);
- *Agriculture and Grasslands*: the remaining cells were assigned to this category.

3.3.3 Land use 4000-3000 BP

Parts of the Meuse catchment were populated by humans at the time of the reference period 4000-3000 BP (Bunnik, 1995; Gotjé et al., 1990; Petit and Lambin, 2002; RWS Limburg/IWACO, 2000). The earliest archaeological evidence of the more or less permanent presence of humans dates from ca. 5300-4800 BC (Venner, 2000). However, any disturbance by humans in this period was, in our opinion, minor, and too small to have had any significant influence on hydrological processes (e.g. Bunnik, 1995; Gotjé et al., 1990; Lefevre et al, 1993; Tallis, 1990). Since the natural vegetation of north-western Europe at that time was predominantly deciduous forest (Adams and Faure, 1997; Bunnik, 1995; Gotjé et al., 1990), it was assumed that the basin was fully forested at 4000-3000 BP, except for cells referring to water bodies and wetlands.

3.4 Water holding capacity maps

A map showing the maximum water holding capacity (WHC) of the soil ($mm.m^{-1}$) is used in STREAM for the calculation of evapotranspiration, direct runoff, groundwater seepage, and baseflow. For the present day the United Nations Food and Agriculture Organization's (FAO) map of maximum WHC (FAO, 2003) was used. Since the FAO dataset has too coarse a resolution (5' x 5') to accurately represent the remaining areas of peat bog in the Meuse basin, the grid cells pertaining to peat bogs were assigned a value of 71 $mm.m^{-1}$, i.e. the modal WHC value for the remaining large intact areas of European peat bog (FAO, 2003). Changes in land use cause changes in soil textural properties, and therefore influence WHC. Hence, a WHC map was made for each century. The FAO WHC map is based on dominant soil unit, component soil units, texture, soil phase, and relief, but does not explicitly consider vegetation. A significant positive correlation was found between WHC and the percentage forest cover between the various soil units (Spearman's Rank, $p=0.049$), and a significant negative correlation between WHC and the percentage cover of agriculture and grasslands (Spearman's rank, $p=0.46$). Hence, the percentage difference between the mean WHC of soils covered by forests and those covered by agriculture and grasslands was calculated, and the resulting change factors were coupled with the land use anomalies over time, to produce WHC maps for each century, similar to Mahe et al. (2005). For cells covered by rendzinas (shallow humus-rich soils over limestone) the original WHC of 63 $mm.m^{-1}$ was retained since the WHC of this soil unit is mainly influenced by parent material (FAO, 2003).

3.5 Maps of initial basin conditions

To obtain realistic initial conditions for soil water storage, groundwater storage, snow-cover, and accumulated potential water loss a perpetual simulation technique (Kleinn et al., 2005) was adopted. Prior to running the simulations the model was allowed to spin-up for a period of 100 years (36,000 iterations) in order to obtain realistic initial conditions.

4 River discharge routing

As STREAM was initially designed to simulate discharge on a monthly time-step (Aerts et al., 1999), it does not contain a routing function to account for the time taken for water entering the system to reach the outflow point. Hence, it is assumed that all of the excess water generating runoff flows through the basin, to the outlet point, in one time-step. Since runoff in the upstream sections of the Meuse takes longer than one day to reach Borgharen, an offline routing component was added in the present study. The routing algorithm is based on the average flow times between specific points on the Meuse (Berger, 1992). In Fig. 7 points are shown for which it takes approximately one day for water to flow between that point and the next downstream point. Daily discharge (Q) at downstream point (p) on day (d) is calculated as:

$$Q_{p(i)d(i)} = (Q'_{p(i)d(i)} - Q'_{p(i+1)d(i-1)}) + (\sum_{i=1}^{n-1}(Q'_{p(i+1)d(i-1)} - Q'_{p(i+2)d(i-2)})) + Q'_{p(n)d(n)} \quad (4)$$

where Q' is the discharge calculated prior to the routing algorithm, n is the total number of flow-days upstream from p until source, and i =1.

Using this approach a simplifying assumption is made that average flow velocities remain the same regardless of discharge totals and flow depths. In reality flow velocities on different reaches of the river change according to the time-specific discharge characteristics. However, for a model designed to compare changes in discharge over millennial time-periods this simplified approach is reasonable, and avoids the need to estimate changes in channel characteristics over the last 4000 years. Moreover, this method was used in a pilot study of the Rhine and Meuse basins, using observed climate data to force the STREAM model for the period 1990-2000, and the model yielded good results (not shown here).

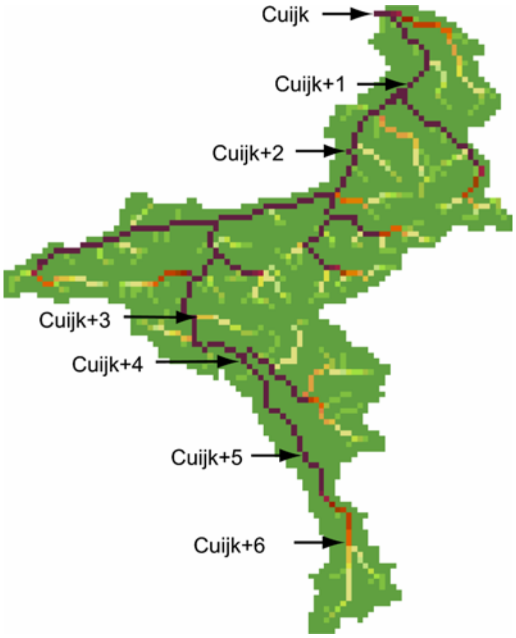

Fig. 7: Points representing k flow-days upstream from Cuijk, e.g. Cuijk+1 is one flow-day upstream from Cuijk, Cuijk+2 is two flow-days upstream from Cuijk, etc. The flow-day points are estimated based on the work of Berger (1992), and are used to route river discharge through the basin on a daily basis.

In order to run STREAM on a daily time-step, a simple threshold-based direct runoff component was added to simulate infiltration excess overland flow. When precipitation exceeds a threshold (Ward and Robinson, 1990), the excess precipitation runs off regardless of whether the soil is saturated or not.

5 Calibration and validation

The calibration of STREAM was carried out by varying model parameters with the aim of reproducing daily, monthly, and annual discharge characteristics similar to those of observed records. A similar approach has been successfully employed in numerous studies (e.g. Aerts et al., 2006; Bouwer et al., 2006; Christensen et al., 2004; Kwadijk, 1993; Van Deursen, 1995, 2000b; Ward et al., 2007; Wood et al., 2002). Where independent measurements exist for output variables other than discharge, these were also used in the calibration; in this case AE and the number of snow-cover days. These data were used in order to reduce to some extent the problems of equifinality of discharge parameter estimation (Kavetski et al., 2006; Kirchner, 2006).

Since the four climate ensemble members have statistically similar monthly frequency distributions (KW-test), for both temperature ($p=0.861$) and precipitation ($p=0.974$), the calibration and validation were based on the ensemble member which showed the closest statistical similarity to the ensemble mean (ensemble member C). Based on the principles of Klemeš (1986) the calibration was carried out for the relatively wet and warm period 1961-2000, and validation for the relatively dry and cool period 1921-1960.

The parameters used for calibration are: crop factor; WHC; TOGW multiplier (determines the proportion of surplus water per grid cell that runs off directly or that seeps to the groundwater); C factor (determines the proportion of groundwater that contributes to baseflow); $SNOW_{tmp}$ (temperature threshold below which precipitation falls as snow and above which snowmelt occurs), and MELT (determines the rate of snowmelt when temperature is above a critical value). The calibration parameters selected for this project were: crop factor, 1.2; WHC, 0.9; TOGW, 0.6; C, 1.25; $SNOW_{tmp}$, 1.0; and MELT, 7/30.

5.1 Discharge

The calibration was initially carried out for the Meuse at Borgharen. The data were provided by Rijskwaterstaat RIZA (Institute for Inland Water Management and Waste Water Treatment), and pertain to the 'undivided Meuse'; they are based on discharge measurements at Borgharen, corrected for canal extractions (Bos, 1993) between Liège and Borgharen (De Wit et al., 2007), and are available since 1911. Further stations on the main river body were obtained for Chooz (Berger, 1992) and Stenay (DIREN Lorraine Bassin Rhin-Meuse). Tributary discharge data were obtained from: Direction Générale des Voies Hydrauliques Region Wallonne (http://voies-hydrauliques.wallonie.be/hydro/annuaireintro.do)for the Chiers (Chauvency-le-Château), Semois (Membre), Lesse (Gendron), Amblève (Martinrive), Ourthe (Tabreux), and Viroin (Treignes); Roer and Overmaas Water Board for the Geul (Meerssen), and Roer (Stah); and Berger (1992) for the Sambre (Namur).

The annual hydrographs for Borgharen (Fig. 8) show good correlation between mean annual and mean monthly modelled and observed discharges for the calibration and validation periods, as well as for both periods taken together. The correlation of total annual discharge was assessed by expressing the mean annual modelled discharge as a percentage of the mean annual observed discharge (%), and the correlation of the means of monthly discharge was assessed using the correlation coefficient, *r*, and the co-efficient of efficiency, N&S (Nash & Sutcliffe, 1970).

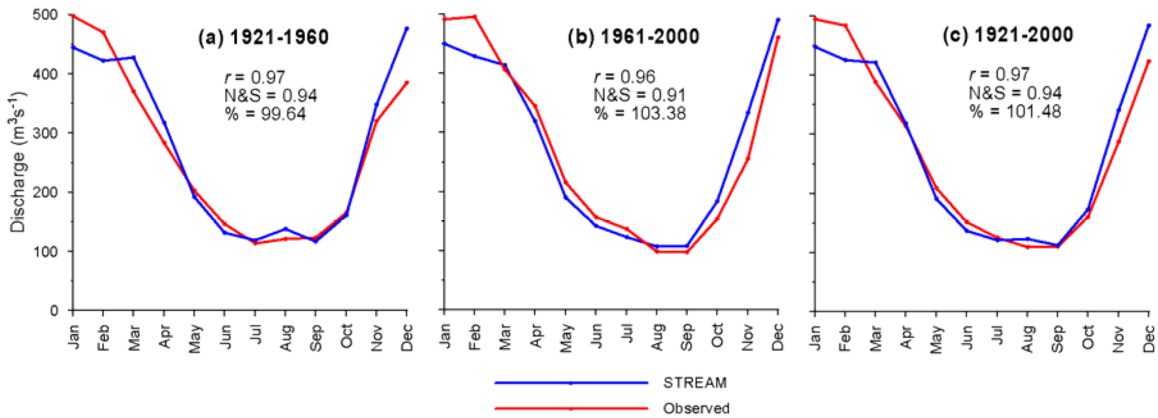

Fig. 8: Annual hydrographs of observed and calibrated model discharge (m^3s^{-1}) at Borgharen for: (a) 1921-1960, (b) 1961-2000, and (c) 1921-2000.

For the upstream gauging stations (Table 2) the correlation is also very good for the Stenay and the Chiers/Semois, and reasonable for the Lesse and Amblève. Although the value of *r* for the Geul is good, the N&S co-efficient, which is also sensitive to absolute differences in discharge, is poor. Hence it is not possible to use this model at the scale of small tributaries.

Table 2: Correlation between modelled and observed mean annual and mean monthly discharge. The calculations were carried out from the earliest year for which the record was available until 2000 inclusive.

Station	River	*Total Accuracy (%)*	*r*	*N&S*
Stenay	Meuse	96.6	0.92	0.84
Chauvency-le-Château/Membre	Chiers/Semois	94.3	0.96	0.92
Gendron	Lesse	76.2	0.98	0.69
Martinrive	Amblève	89.0	0.91	0.68
Meerssen	Geul	129.3	0.92	-1.64

In Table 3 the mean annual k^{th} percentiles of daily discharge are given for the calibration and validation runs (k = 1, 5, 10, 25, 50, 75, 90, 95, 99). The frequency distribution of daily discharge is well calibrated for average and high-flows, although the model slightly overestimates the frequency of summer high-flows (Q_k, k > 97) due to an overestimation of high precipitation events in autumn in ECBilt-CLIO-VECODE. For the entire period 1921-2000 the daily modelled and observed discharge frequency distributions are statistically similar (MWU-test, p=0.366). For very low-flows (Q_1) the agreement between model and observations is good, but a discrepancy exists for Q_5 and Q_{10}.

Table 3: Magnitudes of mean annual k^{th} percentiles of daily discharge (Q_k, k = 1, 5, 10, 25, 50, 75, 90, 95, 99) at Borgharen for the calibration and validation periods. Bold type indicates that the magnitudes are statistically similar to the observed discharge magnitudes (t-test, α=0.05).

	Discharge (m^3s^{-1})								
Time-period	**Q_1**	**Q_5**	**Q_{10}**	**Q_{25}**	**Q_{50}**	**Q_{75}**	**Q_{90}**	**Q_{95}**	**Q_{99}**
Calibration (1961-2000)	**47.5**	53.3	60.8	**93.2**	**185.9**	**380.4**	**607.9**	**754.9**	**1096.2**
Validation (1921-1960)	**46.9**	53.4	61.9	**91.1**	**178.6**	**379.2**	623.1	**774.7**	**1165.7**

For the upstream measuring station at Stenay, and the gauging stations on the Chiers/Semois, Lesse, and Amblève, the daily discharge frequency distributions also show good agreement with the observed series (Fig. 9), despite the relatively short observation periods in some cases.

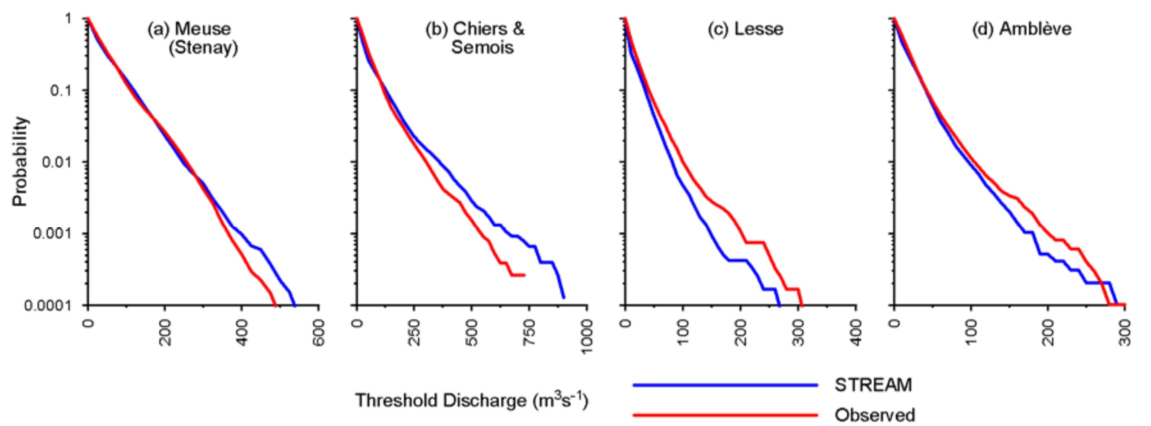

Fig. 9: Logarithmic probability plots of daily discharges over a threshold at: (a) Stenay, (b) Chauvency-le-Château (Chiers) & Membre (Semois), (c) Gendron (Lesse), and (d) Martinrive (Amblève).

5.2 Other calibration parameters

The simulated mean annual basin-average PE for 1921-2000 (605 mm) is of the same order of magnitude as the estimated PE for the Belgian subcatchments for 1968-1998 (555 mm; Leander et al., 2005); the correlation between mean monthly PE totals is good (N&S=0.79, *r*=0.92). Data pertaining to the number of days per year on which the ground is covered by snow at Maastricht (Klein Tank et al., 2002, http://eca.knmi.nl/), were used in the calibration of parameters pertaining to snowmelt and snowfall. Given that the downscaled ECBilt-CLIO-VECODE temperature data slightly underestimate the number of days on which temperature is less than 0°C, the parameter $SNOW_{tmp}$ was set to 1.0°C, which resulted in a good simulation of snow-cover. In the model output for 1961-2000 the ground was covered by snow at Maastricht on average 3.6% of days per year, as opposed to 3.7% of days in the observed record. This parameter could not be examined for the validation period, since the observed data are only available from 1956.

5.3 Sensitivity analysis

To assess the sensitivity of the hydrological model to changes in the calibration parameters, the model was re-run for the calibration period (1961-2000), but altering the values of individual calibration parameters by ±10% and ±50%. For the three calibration parameters which showed the most sensitivity to changes in their values, the results are shown in Table 4. The model is sensitive to changes in crop factor, fairly insensitive to parameters pertaining to WHC and slope, and very insensitive to changes in the other parameters.

Table 4: Results of calibration parameter sensitivity analysis, obtained by running the STREAM model with climate ensemble member C for the period 1961-2000, and adjusting individual calibration parameters by ±10% and ±50%. The correlation between the mean monthly values of each simulation and the observed values is given (r and N&S). The model is sensitive to changes in crop factor, and relatively insensitive to changes in soil water holding capacity (WHC) and slope (C).

Discharge scenario	*r*	N&S	Q_{ann} (m^3s^{-1})	Q_{10} (m^3s^{-1})	Q_{90} (m^3s^{-1})
Borgharen observed	*N/A*	*N/A*	275.1	61.0	625.0
Borgharen simulated	*0.97*	*0.94*	274.1	57.9	613.3
Crop factor - 10%	*0.96*	*0.90*	300.7	70.0	645.9
Crop factor + 10%	*0.97*	*0.92*	251.7	48.7	586.4
WHC - 10%	*0.97*	*0.93*	277.0	59.4	618.4
WHC + 10%	*0.97*	*0.95*	271.5	56.8	608.7
C - 10%	*0.97*	*0.94*	274.1	54.0	617.4
C + 10%	*0.97*	*0.94*	274.1	61.4	610.4
Crop factor - 50%	*0.89*	*-0.57*	445.2	150.3	812.8
Crop factor + 50%	*0.98*	*0.18*	148.9	18.5	406.3
WHC - 50%	*0.95*	*0.88*	292.6	68.3	643.0
WHC + 50%	*0.99*	*0.95*	253.8	50.4	577.0
C = 1	*0.97*	*0.94*	274.1	49.4	622.1
C + 50%	*0.95*	*0.90*	274.3	79.4	597.9

It is imperative that the crop factors used are reliable, as incorrect crop factors could lead to large errors in the estimation of evapotraspiration. Hence, the present day percentage cover of forest and agriculture was calculated for eleven subcatchments (upstream from the following points): Stenay, Chauvency-le-Château, Membre, Gendron, Martinrive, Meerssen, Chooz, Tabreux, Stah, Namur, Treignes. For each subcatchment the percentage anomaly between annual observed and modelled discharge was also calculated. No correlation was found between the percentage forest cover and model accuracy (Spearman's Rank, p=0.467), which suggests that the accuracy of the model is not biased by the crop factor values used for forest and agriculture. Since the main land use changes of the late Holocene have been between these two land use types, the model should therefore be able to simulate the effects of these land use changes on evapotranspiration.

6 Conclusions

A coupled climate-hydrological model was set-up for the simulation of daily Meuse discharge on millennial timescales. The coupled model is made up of a climatic component (ECBilt-CLIO-VECODE) and a hydrological component (STREAM).

The ECBilt-CLIO-VECODE climate data were downscaled to the resolution of the hydrological model. The temperature data show a very good agreement with the observed record in terms of their annual, monthly and daily characteristics. For precipitation the annual and monthly characteristics are well preserved, but the ECBilt-CLIO-VECODE data show too few dry days. However, the frequency and magnitude of high precipitation events has been simulated well.

The hydrological model has been calibrated for the period 1961-2000, and validated the period 1921-1960. The modelled data were compared to independent observed data on discharge for the main Meuse and its tributaries, as well as data on evapotranspiration and snow-cover. In general the modelled calibration results show good agreement with observed annual, monthly and daily discharge characteristics. The frequency of high-flow and very low-flow events appears to be well simulated by the model. Since the model also performs well for the relatively dry and cool validation period (1921-1960), it is assumed that it is well suited to the simulation of the discharge response to moderate climatic changes. However, the simulation of monthly and daily discharges of the Geul tributary was poor, and hence the model should only be applied at the basin-scale or at the scale of major tributaries.

Model accuracy is not affected by the relative percentage cover of forest and agriculture within a subcatchment, and hence the model can be used to examine the sensitivity of Meuse discharge to long-term changes in land use. To facilitate this, a GIS database of land use and soil water holding capacity changes over the periods 4000-3000 BP and 1000-2000 AD has been created. These datasets form a useful resource for other members of the research community interested in the palaeohydrology and palaeoecology of the Meuse basin.

Given that the model approach used here is capable of simulating the effects of changes in climate and land use on Meuse discharge, and can be run efficiently on millennial timescales, it is well suited to the simulation of Meuse palaeodischarge over the periods 4000-3000 BP and 1000-2000 AD, which will form the next phase of this research project.

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8 Appendices

Appendix 1

Equations used in STREAM for the estimation of potential evapotranspiration, actual evapotranspiration, soil water balance and runoff generation.

Potential evapotranspiration

Potential evapotranspiration (ET_{pot}) is based on a modified version of the Thornthwaite (1948) equations, and is defined as:

$$ET_{pot} = \left(\frac{ET_{ref} \cdot CropF \cdot CropF_c}{30} \right) \tag{5}$$

where, if $T \leq 26.5$ then $ET_{ref} = 16\left(10\frac{T}{H}\right)^A$ (6)

or, if $T > 26.5$ then $ET_{ref} = -415.85 + 32.24T - 0.43T^2$ (7)

and $A = 0.49239 + 0.01792H - 0.0000771771H^2 + 0.000000675H^3$ (8)

where $CropF$ is the crop factor, $CropF_c$ is a calibration parameter, T is the mean daily temperature (°C), and H is the HEAT parameter of Thornthwaite (1948), defined by:

$$H = \sum_{jan}^{dec} \left(\frac{T_m}{5}\right)^{1.514} \tag{9}$$

where T_m is the long-term average monthly temperature (°C).

Actual evapotranspiration and the water balance

The actual evapotranspiration is calculated based on the Thornthwaite and Mather (1957) equations:

if $P_{eff} \geq 0$ then $AE = ET_{pot}$ (10)

or, if $P_{eff} < 0$ then $AE = P + MELT - Q_{Hort} + \left(SS_{t^{-1}} - WHC^{\left(-\frac{APWL}{WHC}\right)}\right)$ (11)

and, if $AE > ET_{pot}$ then $AE = ET_{pot}$ (12)

where $P_{eff} = P + MELT - ET_{pot} - Q_{Hort}$ (13)

$SS = SS_{t^{-1}} + P_{eff} - SO$ (14)

$SO = SP + AE$ (15)

and, if $SS \geq WHC$ then $APWL = 0$ (16)

or, if $SS \geq WHC$ then $APWL = WHC \cdot \ln\left(\frac{WHC}{SS_{t^{-1}}}\right) - \left(P + MELT - ET_{pot}\right)$ (17)

where P is the daily rainfall (mm), AE is actual evapotranspiration, $MELT$ is the amount of snowmelt water (mm), $SS_{t^{-1}}$ is soil storage (SS) in the previous iteration (mm), WHC is the soil water holding capacity (mm/m), Q_{Hort} is infiltration excess overland flow (quasi-Hortonian runoff, see below), and SP is soil seepage (see below).

Groundwater balance and discharge generation

The groundwater storage (GW) is calculated as follows:

$$GW_a = GW_{t^{-1}} + SP \tag{18}$$

where, if $P_{eff} \geq 0$ then $SP = (1 - TOGW_c)(SS_{t^{-1}} + P_{eff} - WHC)$, (19)

and, if $P_{eff} < 0$ then $SP = 0$ (20)

$$GW = GW_a - Q_{base} \tag{21}$$

where $GW_{t^{-1}}$ is the groundwater storage in the previous iteration, $TOGW_C$ is a calibration factor which separates between direct runoff and seepage to groundwater, C is a calibration parameter based on cell topography, and Q_{base} is the baseflow, defined as:

$$Q_{base} = \frac{GW_a}{C} \tag{22}$$

Saturation excess overland flow (Q_{over}) is defined as:

if $P_{eff} \geq 0$ then $Q_{over} = TOGW_c \cdot (SS_{t^{-1}} + P_{eff} - WHC)$ (23)

or if $P_{eff} < 0$ then $Q_{over} = 0$ (24)

Infiltration excess overland flow (Q_{Hort}) is defined as:

if $P + MELT > P_{thres}$ then $Q_{Hort} = P - P_{thres}$ (25)

and if $P + MELT \leq P_{thres}$ then $Q_{Hort} = 0$ (26)

Total discharge per grid cell (Q) is defined as:

$$Q = Q_{base} + Q_{over} + Q_{Hort} \tag{27}$$

Appendix 2

Reconstructed Meuse basin land use maps for the late Holocene, showing the large decrease in forested area between 4000-3000 BP and 1000-2000 AD. In the 20th Century AD some reforestation occurred. The maps are based on CORINE data, census data, historical records, and pollen analyses (see Section 3.3).