



Contents lists available at ScienceDirect

Quaternary International

journal homepage: www.elsevier.com/locate/quaint

Last interglacial vegetation in northern Asia: Model simulations and comparison with pollen-based reconstructions

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ARTICLE INFO

Article history:

Available online xxx

ABSTRACT

This study is concerned with vegetation modelling in northern Asia and its comparison with pollen-based vegetation and climate reconstructions. With the help of BIOME1, a global biogeographic equilibrium vegetation model, vegetation dynamics during the last interglacial (LI) period in the northern Asian region can be simulated. In the study, an alternative algorithm to calculate a moisture index (α) is implemented, and the vegetation is modelled for the optimum phase (125 ka BP) and the end of the LI period (115 ka BP). The simulations show an approximately 28% increase in the area occupied by forest biomes for the 125 ka BP time-slice and an approximately 14% decrease in forest-dominated areas around 115 ka BP in comparison to today. Compared to the present day, the tree-line for the 125 ka BP time-slice was located on average $4.05 \pm 2.03^\circ$ further north. Comparison of the modelled results with available pollen-based reconstructions shows that the simulated and reconstructed results for the optimum of the LI period are consistent. With both algorithms calculating the moisture index α , i.e. α_1 and α_2 , the tree line shifts back ($\sim 10 \pm 4.4^\circ$) between the optimum and the end of the LI period. The area occupied by forest in both simulations had declined by an average of $\sim 42\%$. Furthermore, the data-model comparison shows that the modelling results obtained with the newly implemented moisture index are in slightly better agreement with the climate reconstructions.

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1. Introduction

The terrestrial biosphere plays a pivotal role in the Earth's climate through biophysical interactions with the atmosphere and hydrosphere (Foley et al., 1994; Texier et al., 1997). Therefore, understanding the feedback mechanisms between climate and biosphere is a vital key for the description of the past climate and its future development. As with climate, the vegetation can be simulated with numerical models, resulting in a variety of approaches. Because plant growth and vegetation composition are strongly dependent on climate (e.g. Prentice et al., 1992), modelling vegetation changes, especially when compared with vegetation records, has been shown to have a high potential for objective simulation of the natural vegetation and climate history (Kleinen et al., 2011). Furthermore, an excellent overview of the development of dynamic global vegetation models (DGVMs) by several research groups focusing on natural environmental and anthropogenic processes is

given in Prentice et al. (2007), examining 'plant geography', 'plant physiology and biogeochemistry', 'vegetation dynamics', and 'biophysics' as well as 'human intervention: agriculture, urbanisation and forest management'.

Several vegetation models such as BIOME4 (Kaplan et al., 2003), LPJ (Sitch et al., 2003) and ORCHIDEE (Krinner et al., 2005) use various equations to simulate physiological processes such as photosynthesis, carbon allocation and respiration to simulate plant growth. However, as there is no analytical basic equation to describe plant growth as a whole, the simulation of vegetation involves several problems. The present case study uses the BIOME1 model (Prentice et al., 1992), a biogeographic equilibrium-model. This model establishes a connection between five bioclimatic indices and vegetation dynamics. These indices are the mean monthly temperature of the warmest and coldest months (T_w , T_c), the annual sum of growing degree days above 5°C and 0°C (GDD_5 , GDD_0) and a moisture index (α). The moisture index α was introduced in order to describe a plant's moisture budget (Priestley and Taylor, 1972; Wilson et al., 2001; Ajami et al., 2011). The BIOME1 model should be forced with outputs from a climate model at equilibrium. Based on such an equilibrium climate, the spatial

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distribution of ecosystem complexes can be compared with vegetation maps or proxy data. The BIOME1 model was successfully used to simulate palaeovegetation despite its intermediate level of complexity (Prentice et al., 1996; Roeckner et al., 1996; Sánchez Goñi et al., 2005).

For those palaeoclimatic reconstructions and simulations, the last interglacial (LI), the most recent analogue for the current interglacial period with similar climatic conditions (Brewer et al., 2008), is of special interest because the climate and vegetation dynamics during the LI period may provide important insights to enhance our understanding of the Earth's system under the recent progress of global warming (IPCC, 2007). For this period, research has focused on the sea-level and vegetation dynamics as well as the influence of solar radiation (Cuffey and Marshall, 2000; Kubatzki et al., 2000; Crucifix and Loutre, 2002; Cubasch et al., 2006; Schurgers et al., 2006). In general the climatic conditions during the LI had higher temperatures than today (Kukla et al., 2002). The beginning of the LI has been set at ca. 130 ka, and its end at ca. 115 ka (Jouzel et al., 2007). Sánchez Goñi et al. (2005) reconstructed the LI vegetation and climate gradient in Western Europe with the help of pollen data and compared the results with circulation and vegetation model simulations. Kaspar et al. (2005) simulated the climate with a coupled atmosphere ocean model (ECHO-G) for the optimum of the LI (125 ka BP) and used the simulation results for a data-model comparison in the European region. The investigation shows that the model used here (ECHO-G) is able to simulate the climate during the LI optimum satisfactorily. The five climate indices that are needed to drive the BIOME1 model have been calculated by Kageyama et al. (2001) with climatic outputs from different climate models from the PMIP project simulations for Europe and western Siberia. There, MoBidiC (Gallée et al., 1991) was used to model the atmosphere whereas BIOME1 was used to compute the moisture index. The simulations were done for the last glacial maximum, ca. 21 ka BP, and showed a broad agreement with pollen-based reconstructions, except for Western Europe where the moisture index α is overestimated by the model simulations.

In this study, we compare results of model simulations to climate and vegetation reconstructions based on proxy records which may help to validate these results to better understand the different climate factors influencing past environments within the region of north Asia. Until now, there have been only a few investigations (Harrison et al., 1995; Melles et al., 2012) on this subject for the northern Asian region during the LI period. This raises the question whether the mentioned discrepancy between model simulations and proxy reconstructions is only valid for glacial conditions or also for interglacial phases in this region. To answer this question, we first examine the vegetation dynamics during two pronounced phases of the LI – at the optimum (125 ka BP) and end (115 ka BP) of the LI (Jouzel et al., 2007). The BIOME1 model has been used to simulate the vegetation, whereas the vegetation is calculated using climate simulations according to Kaspar et al. (2005). Furthermore, we implement an alternative method for calculating α and investigate whether this provides an improvement compared to the original method described by Prentice et al. (1991). Furthermore, the obtained simulation results are compared with published pollen-based vegetation and climate reconstructions from sites in northern Asia.

2. Experimental design

2.1. Calculation of moisture index α

In this study, two approaches of calculating the moisture index α (further called α_1 and α_2 , respectively) are presented. Therefore, two different sets of physical climate variables are used to drive the

BIOME1 model. The first approach calculates α_1 (moisture index as in Prentice et al., 1992) using total cloud cover, surface temperature, total precipitation and available water capacity. Alternatively, in the current study moisture index α_2 is calculated differently from Prentice et al. (1992) using surface temperature, and latent and sensible heat flux.

The vegetation cover is described by a combination of so called plant functional types (PFTs) whereas the PFTs are defined by several environmental and ecological properties (Woodward and Cramer, 1996); these are basic physiognomy (trees and non-trees) and leaf characteristics (evergreen and deciduous, broad-leafed and needle-leafed). In BIOME1 thirteen PFTs and an additional dummy type are defined and allocated to climatic conditions according to the climate variables of T_c , T_w , GDD_5 , GDD_0 , and α (Prentice et al., 1992). The latter variable can be considered as an integrated measure of the annual amount of growth-limiting drought stress on plants (Prentice et al., 1992). The index is also called the Priestley–Taylor coefficient because it is the linear constant of proportionality in the Priestley–Taylor equation $E = \alpha^* E_q$ (Priestley and Taylor, 1972). E is the evapotranspiration and E_q the equilibrium evapotranspiration, which is an expression for the radiation part within the Penman–Monteith equation for evapotranspiration (Monteith, 1965). All variables besides α can be derived directly from the climate model output. In the vegetation simulation, the plant competition for sunlight and soil moisture is expressed with a dominance class (D). After applying the environmental limits to determine which plant types can occur, the dominance class values of these types are examined and only those in the highest class (lowest D) are retained in the code indicating the biome type. The climate indices are calculated on a yearly basis and an average of the whole simulation period is applied to assign the dominance type of the PFTs. In the second step, the PFTs are assigned to their corresponding biomes (Prentice et al., 1992).

The procedure applied in this study can basically be separated into two steps. First, the new implemented algorithm for calculating α is tested with NCEP/NCAR reanalysis climate data (NOAA and NCAR, 1949–2011; Kalnay et al., 1996) in order to compare the results with a satellite based vegetation map from Olson et al. (1983). Secondly, the vegetation during the optimum and end of the LI is simulated with the help of ECHO-G simulations (Kaspar et al., 2007) and tested with the Cohens- κ -statistics (Cohen, 1968; Monserud and Leemans, 1992). All reanalysis data is provided by the physical-scientific department of NOAA-ESRL in Boulder, Colorado via the website <http://www.esrl.noaa.gov/psd/>. The grid-resolution of the modern climate data is T62 (ca. 1.875°). The BIOME1 model is driven by an equilibrium climate based on the years from 1978 to 2008. As this work focuses on the simulation of palaeovegetation derived from a climate model, the validation is based on reanalysis climate data and not on observational data.

2.2. Last interglacial (LI) vegetation simulation

The equilibrium climate for the palaeovegetation simulation is based on ECHO-G climate specifications (Legutke and Maier-Reimer, 1999; Legutke and Voss, 1999). The ECHO-G is a coupled atmosphere ocean climate model, for which the ECHAM4 is the basis for the atmosphere simulations (Roeckner et al., 1996). The spatial horizontal resolution of the atmosphere model is T31 ($\approx 3.75^\circ$). The model has 19 vertical hybrid sigma-pressure levels. The basis for the ocean component of ECHO-G is HOPE-G (Wolff et al., 1997). It provides a dynamic-thermodynamic sea-ice model (Hibler, 1979) with a spatial resolution of T42 ($\approx 2.8^\circ$) and 20 vertical, irregularly ordered layers. ECHO-G control run simulations by Min et al. (2005a) show that the present climatology and inter-annual variability of surface temperature, precipitation and sea

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