The Terrestrial NPP Simulation in China since 6ka BP *

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ABSTRACT

A better understanding of the long-term global carbon cycle required estimate of the changes in terrestrial carbon storage after the last glacial period. The results of simulation at mid-Holocene (MH) from PMIP (Paleoclimate Modeling Intercomparison Project) and the modern data from CRU (Climate Research Unit, East Anglia University, UK) allow us to use the Atmosphere-Vegetation Interaction Model (AVIM) to simulate the Chinese terrestrial net primary productivity (NPP) at 6ka BP and present time. The change of NPP and total NPP in China from now to mid-Holocene are about 54 g $m^{-2}yr^{-1}$ and 0.63 Pg yr^{-1}, respectively, mainly due to the build-up of temperate forest and tropical rainforest. Chinese terrestrial NPP variation from MH to now is closely related to the variation in intensity of Asian monsoon, which controlled the climate-vegetation pattern change.

Key words: NPP (net primary productivity), mid-Holocene, climate change, Asian monsoon

1. Introduction

According to the reconstruction of paleo-temperature based on δ^{18} O data of ice core in the Greenland (see Jouzel et al., 1987; Grootes et al., 1993; Blunier and Brook, 2001), the current interglacial epoch, the Holocene, began at ca. 11.5 thousand years before present (ka BP). Multiple sources (pollen data, macrofossils) reveal that the summer climate in the Northern Hemisphere was warmer in the early to middle Holocene (MH) (ca. 8-6ka BP) relative to the present climate. This warm has been explained by the variations in the orbital forcing: the Northern Hemisphere received more solar radiation in the boreal summer during the MH than at present (see Berger, 1996). The orbital forcing hypothesis for the MH has been successfully tested with climate models of different complexities (see Kutzbach et al., 1996).

Moreover, ice core data from the Vostok (see Barnola et al., 1993) and the Taylor Dome (see Indermühle et al., 1999) show that the atmospheric CO_2 concentration at MH is about 20 ppmv lower than the preindustrial level. Analyzing the record from the Taylor Dome reveals that an increase in the atmospheric CO_2 concentration is mainly caused by the decay of the terrestrial carbon accumulated in the early Holocene (see Indermühle et al., 1999). Terrestrial biosphere plays an important role in the global carbon cycle as a modulator of atmospheric CO_2 changes. Concerns about the impact of increasing atmospheric CO_2 concentration and future global warming on the terrestrial biosphere, as well as the interactions and feedbacks between the climate system and the terrestrial biosphere, have focused attention on our need to better understand the past changes in terrestrial vegetation and carbon storage.

The mainland of China, located at 20°-54°N, 75°-130°E, has climate regimes ranging from perennial snow on the western high mountains to deserts in the western lowlands, and from cold temperate regions in the northeast to warm and humid tropics along the southern coast. The climate of east-central China is characterized by alternations of East Asian summer monsoon and winter monsoon from Siberia, driven by the differential heating between the Asian Continent and the Pacific Ocean to the east and southeast, and

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the Indian Ocean to the southwest. The monsoons support a unique set of ecosystems ranging from the boreal coniferous forest in the northeast to the tropical rainforest in the south. The climatic variability, topographic complexities, and natural ecosystem diversity give China an important role in global change studies (see Ye et al., 1995).

Great efforts have been made to reconstruct the changes in paloeclimate patterns and paleovegetation distributions, however the regional-scale reconstruction of long-term terrestrial carbon of China has less been focused on (see Peng, 1997). The objective of the present study is to use the AVIM (Atmosphere-Vegetation Interaction Model) to investigate the terrestrial carbon cycle in China since MH, and our main attention is on the net primary productivity (NPP) of the vegetation at mid-Holocene and present time. The NPP distribution and its relationship with the climatic change have also been explored.

2. Model and data

2.1 *Description of the atmosphere-vegetation interaction model*

AVIM (Atmosphere-Vegetation Interaction Model) is a climate-vegetation interaction model, which has in two ways coupled the land surface physical process with the eco-physiological process, and achieved the dynamic interaction between atmosphere and terrestrial vegetation (see Ji, 1995). Validations of AVIM have been conducted in forests, croplands, and grasslands of China, and in global ecosystems (see Ji 1995; Li and Ji, 2001). The model has participated in the Ecosystem Model-Data Intercomparison (EMDI), and was proved to have good performance of NPP simulation. There are three subcomponents in this model: physical exchange module, vegetation growth module, and physical parameter transformation module. The physical processes related to energy and water cycle and the vegetation growth processes related to carbon cycle are included in AVIM. Main processes of the vegetation are photosynthesis, respiration, falling of leaf, stem and litter decomposition, and the changes in biomass for each tissue are determined by the budget of carbon. The photosynthesis process related to NPP of terrestrial vegetation is controlled by temperature, soil moisture, and CO_2 concentration in the atmosphere (see Ji, 1995), the NPP calculated by the AVIM can be written as

$$\mathrm{NPP} = A - R, \tag{1}$$

where A is the minimum carboxylation rate following the limitations of Rubisco, RuBP, and TUP (see Farquhar et al., 1991; Richardon, 1981), and controlled by environmental factors. R is the respiration rate. We have

$$A = F[f(C_a), f(T_f), f(\varphi)] \tag{2}$$

where C_a is the CO_2 concentration in the atmosphere, T_f is the canopy temperature, and ψ is the water potential that is related to soil moisture.

2.2 *Data*

In this study, the time step of AVIM is 30 min, and the spatial resolution is $0.5° \times 0.5°$ (longitude-latitude grid cell). Modern monthly mean climate data with $0.5° \times 0.5°$ resolution are downloaded from CRU (Climate Research Unit, East Anglia University, UK), and the MH climate data come from monthly mean datasets of PMIP (Paleoclimate Modeling Intercomparison Project). We choose the simulation results of the ECHAM3 and UGAMAP for Chinese terrestrial NPP simulation at 6ka BP in our research. The annual mean temperatures from UGAMP and ECHAM3 increase by 2-4°C and 6°C more than that on the Tibetan Plateau (Table 1). Meanwhile the annual precipitation increases by even more than 100% compared with the present time. The above simulations of PMIP in China are consistent with the geological data and other paleo-climate simulations (see Shi et al., 1993; Chen et al., 2002), thus are reliable as initial climate

Table 1. The comparison of annual mean temperature (AMT) and annual total precipitation (ATP) in China at 6ka BP and present time

Time	Model	AMT (°C)	ATP (mm)
6ka BP	ECHAM3	9.16	843
	UGAMP	8.37	960
Now	CRU	5.35	626

forcing to run AVIM.

The CO_2 concentrations of the atmosphere at the present and MH were set to 345 ppm and 270 ppm, respectively, suggested by PMIP. In this paper we neglected the sea level effect and suggest the Chinese areas at the two eras are the same. The modern vegetation dataset in China is classified with the Dorman and Seller standard (Dorman and Sellers, 1989), vegetation types displayed include: (1) tropical rainforest, (2) broad leaf deciduous trees, (3) broad leaf and needle leaf trees, (4) needle leaf evergreen trees, (5) needle leaf deciduous trees, (6) broadleaf trees with ground cover, (7) grassland, (9) sparse shrub, and (11) crops. The vegetation distribution map of MH is based on the research of Shi (1993), digitized with the $0.5° \times 0.5°$ grid cell under the same standard, which contains six types of vegetation. Comparing this map with modern vegetation distribution, we found that the needle deciduous tree has disappeared and the tropical rainforest and broadleaf deciduous tree increased at MH, as shown in Fig. 1.

The modern soil texture data are classified into six soil types (Zobler, 1986). Because of the lack of soil information for the MH, we simply used the same soil map for the present time.

3. Results

3.1 *Modern distribution of terrestrial NPP in China*

Table 2 shows that modern terrestrial average NPP in China is estimated at 355 g $m^{-2}yr^{-1}$, the highest average NPP is tropical rainforest with 926 g m^{-2} yr^{-1}, and the lowest average NPP is sparse shrub with 159 gm m^{-2} yr^{-1}. We present the other results of simulations and measurements for Chinese vegetation NPP values in Table 3 and the comparison of the NPP values suggests that our results of simulation are feasible. Figure 2 exhibits the terrestrial NPP distribution in China at present time, green color area (NPP: 500 g m^{-2} yr^{-1}) occupies the half in the East China, the highest NPP appears at Xishuangbanna, Yunnan Povince, the tropical rainforest area, southwest of China. The lowest NPP (orange color) appears in Tibet and Xinjiang, west part of China, mainly occupied by the desert and grassland.

Multiplying the NPP of each grid by the corresponding grid area, the total terrestrial NPP of China is calculated, which reflects the amount of carbon absorbed from the atmosphere by vegetation. The total terrestrial NPP in China is 3.33 Pg yr^{-1} at present time, and the highest total NPP values for the different type vegetation are at the grassland and the broadleaf trees with ground cover, both being 0.85 Pg yr^{-1}, due to their high average NPP or large area respectively (Table 2). The lowest total NPP values are at the tropical rainforest and the needle leaf evergreen trees, owing to their less area in China.

3.2 *Change in NPP from MH to the present time*

As shown in Fig.3 and Table 4, average NPP at

Table 2. Reconstruction of present vegetation distribution and the total NPP and average NPP calculated using the AVIM model

Vegetation code	Vegetation type	Grid cells	Average NPP (g m^{-2} yr^{-1})	Total NPP (Pg yr^{-1})
1	Tropical rainforest	34	926	0.08
2	Broad leaf deciduous trees	184	530	0.22
3	Broad leaf and needle leaf trees	191	434	0.21
4	Needle leaf evergreen trees	187	499	0.19
5	Needle leaf deciduous trees	44	354	0.02
6	Broadleaf trees with ground cover	469	668	0.85
7	Grassland	1175	237	0.85
9	Sparse shrub	382	159	0.15
11	Bare soil	589	0	0
12	Crop	586	429	0.76
13	Water	4	0	0
Total		3845	355	3.33

Table 3. Comparison of Chinese vegetation NPP values between AVIM and other resources (units:g·$m^{-2}yr^{-1}$)

Vegetation types	AVIM	CEVSA (Tao et al., 2003)	Sun (2001)	Liu*	Measurement (Liu, 1997)
Needle leaf evergreen trees	499	486	529	587	160-680
Needle leaf deciduous trees	354	345	420	585	150-500
Broad leaf deciduous trees	530	624	460	928	250-700
Broad leaf and needle leaf trees	434	423	403	870	250-1000
Broadleaf trees with ground cover	668	648			
Grassland	237	348		271	
Crop	429	606		752	

*Liu M. L., 2001: Land-use/cover change and terrestrial ecosystem phytomass carbon pool and production in China. Beijing: Institute of Remote Sensing Application of CAS.

Table 4. Reconstruction of the vegetation distribution and the total NPP and average NPP calculated using the AVIM model at MH

Code	Vegetation type	Grid cells	Average NPP(g $m^{-2}yr^{-1}$)	Total NPP(Pg yr^{-1})
1	Tropical rainforest	233	949/913 (931)	0.61/0.63 (0.62)
2	Broad leaf deciduous trees	923	550/536 (543)	1.16/1.11 (1.14)
3	Broad leaf and needle leaf trees	115	426/446 (436)	0.10/0.09 (0.9)
4	Needle leaf evergreen trees	38	476/492 (484)	0.03/0.03 (0.03)
6	Broadleaf trees with ground cover	554	685/722 (704)	1.08/1.02 (1.05)
7	Grassland	1525	233/246 (255)	1.03/0.87 (0.95)
11	Bare soil	457	0/0 (0)	0/0
Total		3845	400/417 (409)	4.04/3.87 (3.96)

Note that the left of slash is the result of UGAMP, and the right is the ECHAM3, and the average of both models is in the brocket.

MH is 409 g $m^{-2}yr^{-1}$, 19.5% higher than that of present. The NPP results of every type vegetatio with the ECHAM3 and UGAMP climate data are nearly the same. The highest NPP values at MH are tropical rainforest with 931 g $m^{-2}yr^{-1}$ and the broadleaf trees with ground cover with 704 g $m^{-2}yr^{-1}$, and the lowest is the grassland with 255 g $m^{-2}yr^{-1}$. The distribution of NPP at MH is similar to now, but the green color area (high NPP value) is larger in the East China and the orange color (low NPP value) area is lower in Tibet comparing with present time, the highest NPP also occurs at Xishuangbanna area, Yunnan Province. For the total NPP it is about 3.96 Pg yr^{-1}, 18.9% higher than that of present, and the highest total NPP values are from the broad leaf deciduous trees being 1.14 Pg yr^{-1} and broadleaf trees with ground cover being 1.05 Pg yr^{-1}, the lowest is needle leaf evergreen only with 0.03 Pg yr^{-1}.

4. Discussions

Estimation of terrestrial NPP for the selected past period should help us better evaluate the role of the terrestrial biosphere in the long-term global carbon scale, as well as quantify and separate changes due to natural variability from those due to human activities. Many efforts have been exerted to reconstruct the global and regional NPP at MH. Globally, Foley (1994) used DEMETER model to estimate differences in terrestrial carbon storage associated with the modern and mid-Holocene climate simulated by the GENESIS global climate model. He found that the global average NPP and total NPP changed less than 5% over last 6000 years BP. However the possible vegetation and soil feedbacks were not considered in the climate simulation of GENESIS, which likely underestimated the increases in summer precipitation by an enhanced frican summer monsoon (Kutzbach et al., 1996). On the other hand, François (1999) suggested that if the CO_2 fertilization is not taken into account the terrestrial total NPP would increase by 0.75-2.26 Pg yr^{-1} from 6ka BP to now, but if the CO_2 is considered it would yield a decrease (-2.7 to -0.98 Pg yr^{-1}) from mid-Holocene to the present, based on the CARAIB model. It is somewhat contradiction on the above results. In Europe Peng et al. (1995) revealed that the terrestrial total NPP increased about 7% from 6ka BP to now using an empirical model. Monserud (1995) found that about 10% (0.6 Pg yr^{-1}) increase of the total NPP of Siberian ecosystem occurred at 6ka BP comparing with present time, mainly due to the shift from dark-needled taiga in the mid-Holocene to light-needled taiga today.

In this work from MH to now the terrestrial average NPP and total NPP in China change -54 g $m^{-2}yr^{-1}$ and -0.63 Pg yr^{-1}, about 15% and 19% decrease respectively. Our results are in agreement with the estimate of past terrestrial carbon storage of China, reported by Peng (1997), who referred that the shifts of terrestrial carbon storage in vegetation were from 70.6 Pg yr^{-1} to 57.9 Pg yr^{-1}, about 22% decrease. The similarity of this region may be associated with effects on paleoclimate and vegetation dynamics by monsoon system variation from MH to now. It is important to note that the variation of terrestrial NPP in China from mid-Holocene to now is closely related to the variations in intensity of Asian monsoons which controlled the changes in patterns of climate and vegetation (An et al., 1991; Xiao et al., 1995). In fact the reconstruction of vegetation distribution at MH shows that the forest moved to north and west mainly due to the northward and westward expansion of the summer monsoon (Shi et al., 1993). The GCM experiments for 6ka BP also show that the strengthening of East Asian monsoon resulted in more precipitation in the East China (Chen et al., 2002). The paleomonsoon-driven patterns support the results of global climate model experiments that predict an enhanced monsoon system between 12 000 and 6000 years BP, followed by a weak monsoon system from 6000 years BP to now (Kutzbach and Guetter, 1986; COHMAP members, 1988). An et al. (1991) also show that the edge of East Asian summer monsoon at MH stayed northern comparing with now.

Here we divide China into two parts along 100°E line as East China, mainly affected by the Asian monsoon, and West China, which is not affected by the Asian monsoon, then calculated the NPP of these two parts in order to estimate the enhanced Asian monsoon effects on the Chinese terrestrial NPP at 6ka BP. Table 5 shows that in East China the total NPP increases by 0.3 Pg yr^{-1} from now to 6ka BP, but in the west it increases less than 0.22 Pg yr^{-1}, and for the average NPP increase are 60 g $m^{-2}yr^{-1}$ and 50 g $m^{-2}yr^{-1}$ respectively for East and West China. The NPP comparison of East and West China means that at enhanced Asian monsoon area (East China) the terrestrial NPP increases are higher than that unaffected area (West China) at MH.

Table 5. The NPP characteristics in eastern and western China at 6ka BP and present time

Time	Average NPP ($g\ m^{-2}yr^{-1}$)		Total NPP (Pg yr^{-1})	
	East	West	East	West
Now	493	153	2.81	0.52
6ka BP	550	178	3.16	0.71
	557	202	3.2	0.76

There are still some uncertainties, however, due to several possible sources of error in this reconstruction. The first source of error is the difficulty of accurately reconstructing the paleovegetation distribution from the sparse paleodata. More accurate paleovegetation maps would help us improve our results, but such maps are not available or published. The second source is the climate data. We used the weather generator to produce the daily data from the monthly, but some uncertainties occurred at this process, and the better way is to run the AVIM coupled with GCM to calculate the NPP, which is our following work. The third possible source of error is the lack of systematic soil information for the MH, and we have no choice but use the modern data. Despite all the possible sources of errors discussed above, we conclude from the reconstruction presented here that the changes in climate and in atmospheric CO_2 resulted in an decrease in the average and total NPP of about 54 g $m^{-2}yr^{-1}$ and 0.54 Pg yr^{-1} respectively from MH to present. Our results clearly show that the monsoon-driven changes in climate and shift in paleovegetation at 6ka BP had an important effect on the long-term carbon dynamics of China, which strongly contributed to the global carbon cycle.

REFERENCES

An Zhisheng, Wu Xihao, Lu Yu, et al., 1991: A preliminary study of the paleoenvironment changes of China during the last 20 000 years. In: *Quaternary Geology and Global Change*. Liu Dongsheng, Eds., Beijing: Science Press, 1-26. (in Chinese)

Barnola J. M., D. Raynaud, C. Lorius, et al., 1993: Vostok ice core atmosphere CO_2 concentration record. In *Trends '93 Database*, CDIAC, ORNL, Oak Ridge, Tenn.

Berger A., 1996: Orbital variations. In: *Encyclopedia of Climate and Weather*. S. Schneider, Eds., New York: Oxford Univ. Press, 557-564.

Blunier T., and E. J. Brook, 2001: Timing of millennium-scale climate change in Antarctica and Greenland during the last glacial period. *Science*, **291**, 109-112.

Chen Xin, Yu Ge, and Liu Jian, 2002: The mid-Holocene climate simulation on East Asia and the discussion on the temperature variation mechanism. *Science in China (Series D)*, **4**, 335-345. (in Chinese)

COHMAP Members, 1988: Climatic changes of the last 18 000 years: Observations and model simulations. *Science*, **241**, 1043-1052.

Crucifix M., M. F. Loutre, P., Tulkens et al., 2002: Climate evolution during the Holocene: A study with an Earth system model of intermediate complexity. *Climate Dynamics*, **19**, 43-60.

Dorman J. L., and P. J. Sellers, 1989: A global climatology of albedo, roughness length and stomatal resistance for atmospheric general circulation model as represented by the simple biosphere. *Journal of Applied Meterology*, **28**, 833-855.

Farquhar G. D., S. V. Caemmerer, and J. A. Berry, 1980: A biochemical model of photosynthetic CO_2 assimilation leaves of C_3 species. *Planta*, 149, 79-90.

Foley J. A., 1994: The sensitivity of ecosystems to climatic change: Simulating the terrestrial biosphere of the middle-Holocene. *Global Biogeochemical Cycle*, **8**, 505-525.

François L. M., Y. Godderis, and P. Warnant, 1999: Carbon stocks and isotopic budgets of the terrestrial biosphere at mid-Holocene and last glacial maximum times. *Chemical Geology*, **159**, 163-189.

Grootes P. M., M. Stuiver, J. W. C. White et al., 1993: Comparison of oxygen isotope records from GISP2 and GRIP Greenland ice cores. *Nature*, **366**, 552-554.

Indermühle A., et al., 1999: Holocene carbon cycle dynamics based on CO_2 trapped in ice at Taylor Dome, Antarctica. *Nature*, **398**, 121-126.

Ji Jinjun, 1995: A climate-vegetation interaction model: simulating physical and biological processes at the surface. *Journal of Biogeography*, **22**, 445-451.

Jouzel J. C., J. R. Lorius, C. Petit et al., 1987: Vostok ice core: Continuous isotope temperature record over the last climatic cycle (160 000years). *Nature*, **329**, 403-408.

Kutzbach J., and P. J. Guetter, 1986: The influence of changing orbital parameters and surface boundary conditions on climate simulations for the past 18 000 years. *Journal of Atmosphere Science*, **43**, 1726-1759.

Kutzbach J. E., G. Bonan, Foley J. et al., 1996: Vegetation and soil feedbacks on the response of the African monsoon to orbital forcing in the early to middle Holocene. *Nature*, **384**, 623-626.

Li Yinpeng and Ji Jinjun, 2001: Simulations of carbon exchange between global terrestrial ecosystems and the atmosphere. *Acta Geographica Sinaca*, **56**, 379-389. (in Chinese)

Liu Shirong, Xu Deying, Wang Bin et al., 1997: The impacts of climates change on productivity of the forests in China. *A Study on the Impacts of Climate Change on Forests in China, In: Xu Deying Eds.* Beijing: China Science and Technology Press, 75-93. (in Chinese)

Monserud R. A., O. V. Denissenko, P. K. Tatyana et al., 1995: Change in phytomass and net primary productivity for Siberia from the mid-Holocene to the present. *Global Biogeochemical Cycles*, **9**, 213-226.

Peng Changhui, J. Guiot, E. Van Campo et al., 1995: Temporal and spatial variations of terrestrial biomes and carbon storage since 13 000 yr BP in Europe: Reconstruction from pollen data and statistical models. *Water, Air and Soil Pollution*, **82**, 375-391.

Peng Changhui, and J. A. Michael, 1997: Contribution of China to the global carbon cycle since the last glacial maximum. *Tellus*, **48**(B), 393-408.

Richardson C. W., 1981: Stochastic simulation of daily precipitation, temperature, and solar radiation. *Water Resources Research*, **17**,182-190.

Shi Yafeng, Kong Zhaocheng, Wang Sumin et al., 1993: Mid-Holocene climate and environment in China. *Global and Planetary Change*, **7**, 219-233.

Sun Rui and Zhu Qijiang, 2001: Effect of climate change of terrestrial net primary productivity in China. *Journal of Remote Sensing*, **5**, 58-61. (in Chinese)

Tao Bo, Li Kerang, Shao Xuemei et al., 2003: Temporal and spatial patterns of net primary production of terrestrial ecosystems in China. *Acta Geographica Sinaca*, **58**, 372-380. (in Chinese)

Xiao Jule, S. C. Porter, An Zhisheng et al., 1995: Grain size of quartz as an indicator of winter monsoon strength on the Loess Plateau of central China during the last 130 000 years. *Quaternary Research*, **43**, 22-29.

Ye Duzheng et al., 1995: China contribution to global change studies. *China Global Change Report No.2*, Beijing: Science Press, 1-226. (in Chinese)

Zobler, L., 1986: A world soil file for global climate modeling. *NASA Tech. Memo.* 87802. NASA Goddard Institute for Space Studies, New York, U S A, 33.