Evidence of Climate Mitigation Feedbacks by Global Forest Cover Expansion during the Pliocene

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Abstract

During the Pliocene, global climate was believed to be in a global greenhouse state, with high mean temperature and rainfall favouring high-latitude forestation, along with a possible high ocean primary productivity and more efficient burial of organic carbon. These phenomena could have increased the quantitative role of CO2-related negative feedbacks, leading to a gradual damping effect on the Pliocene orbitally-paced climate responses, as documented by the asymptotic decay (up to -100%) of the global $\delta^{18}O$ and equatorial sea surface temperature response sensitivity.

Keywords: Forestation; Deforestation; Marine phytoplankton; Climate mitigation; Negative feedback; Damping mechanisms; Solubility pump; Biological pump; Albedo

Introduction

Deforestation refers to the global decrease in forest area arising from natural or anthropogenic causes. Since 1960, human activities have been among the main causes of deforestation. According to the Food and Agriculture Organization, the expansion of agriculture caused nearly 80% of global deforestation, together with mining activities and urbanisation [1]. Deforestation has been negatively affecting natural ecosystems and biodiversity, but the influence of vegetation changes on climate remains unclear. The overall impact of forestation on climate depends on carbon uptake by photosynthesis, which leads to reduced radiative forcing (cooling), but also on other effects associated with the development of new forests, particularly the decrease in albedo, which exerts a positive radiative forcing (warming) on climate [2,3]. The net effect depends on the relative magnitude of these two opposing processes, which can be calculated, but with considerable margins of uncertainty on the basis of the assumptions made, and this leads to apparently conflicting results [3]. In many boreal forest regions, the positive forcing induced by decreases in albedo can offset the negative forcing that is expected from increased forest carbon sequestration [2]. In contrast, in quite a different study which introduced an interesting *time factor* under New Zealand conditions, Kirschbaum et al. [3] reported that forest albedo and carbon-storage effects were of similar magnitude for the first four to five years after tree planting, but as the stand aged, carbon sequestration increasingly dominated,

giving a net cooling effect. Excluding the time factor, the main differences between these findings from different regions relate to the role of snow (snow plays an important role in the boreal zone, but only a minor one in New Zealand), the growth rate of trees and their carbon-storage potential (much greater in New Zealand than in more slowly growing boreal forests), and annual incident solar radiation (greater in New Zealand than in the boreal zone) [2,3]. These studies show that the regional effects of forest cover may be conflicting, but we must consider that the global result is clearly a net balance of regional effects that take time. This introduces another important component that should be considered in the overall balance of climate mitigation effects, namely, the role of marine phytoplankton as a biological pump [4-7].

This short paper aims to highlight an interesting recent observation on a global climate mitigating effect from the geological record to stimulate the scientific community to further explore this important issue and thereby increase our understanding of the dynamics of the climate system and its damping mechanisms.

Asymptotic Decay of the $\delta^{18}O$ and SST response Sensitivity in the Pliocene

Figure 1 shows a quantitative analysis of the Plio-Pleistocene global LR04 $\delta^{18}O$ [8] and equatorial ODP Site 846 sea surface temperature (SST) [9] response sensitivity, relative to nominal orbital forcing (eccentricity La2010, [10]; obliquity and precession

La2004, [11]), as a function of the long-term mean climate state, from the work of Viaggi [12]. Benthic $\delta^{18}O$ stack measures changes in global ice-volume and deep-water temperatures, which are controlled by high-latitude surface temperatures [13]. The response sensitivity R_s is given by the following function [12]:

$$R_s = \left(s^2_{resp} - s^2_{forc} \right) / s^2_{forc} * 100$$

where s^2_{resp} and s^2_{forc} are the variance of the standardised (0-mean; 1-standard deviation) orbital response component and nominal forcing (eccentricity, obliquity, precession) calculated by time segments binned at 532kyr. A self-sustained climate system, which is only paced by orbital cycles, would produce a response sensitivity near zero because the variance of the response would match the variance in forcing, whereas values much larger than zero indicate a non-linear reinforcement of the signal response. Negative R_s reflects a response variance lower than orbital forcing, suggesting a damping mechanism for the climate system.

Figure 1 exhibits a remarkable, gradual asymptotic decay of both global $\delta^{18}O$ and equatorial SST R_s towards the Pliocene which all become negative, to -100%, suggesting a damped orbital response during the time of minimum average ice volume and higher mean temperature.
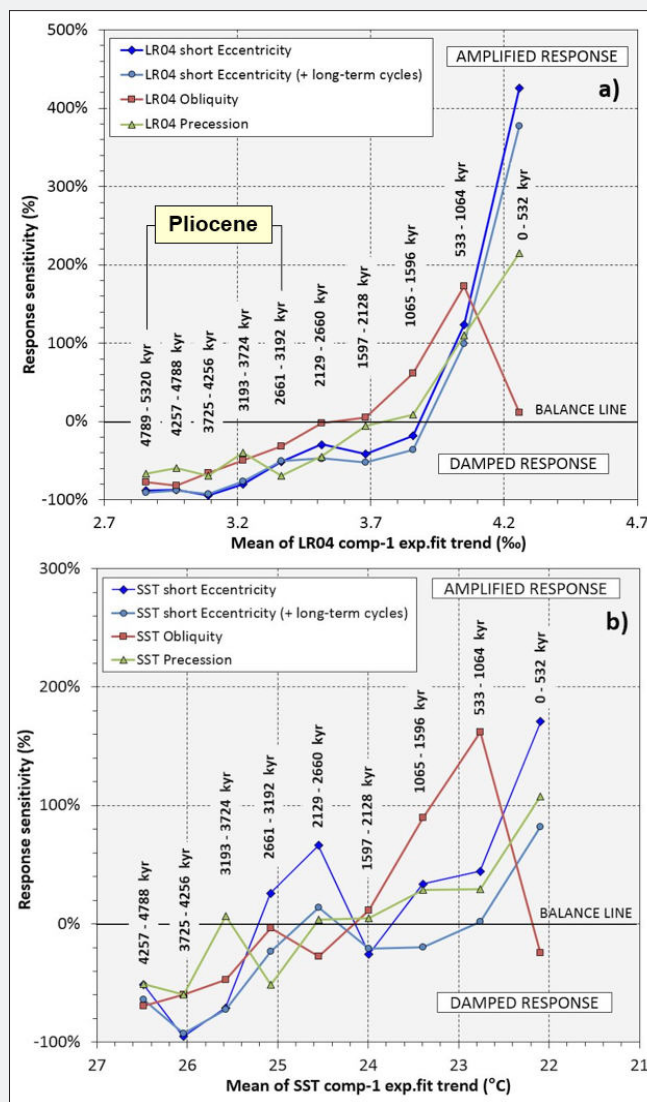

Figure 1: Analysis of the magnitude, during the Pliocene-Pleistocene, of (a) global δ18O and (b) equatorial SST response sensitivity (Rs) relative to orbital forcing as a function of the long-term mean climate state by time segments binned at 532kyr. The balance line (0%) indicates that the variance of response matches the variance in forcing, suggesting a self-sustained climate system, which is only paced by orbital cycles. Rs becomes negative in the Pliocene and indicates a response variance lower than orbital forcing, suggesting a damping mechanism for the climate system. Note the strong non-linear asymmetry between amplified responses, to 400% global δ18O, and damped responses, to -100% (modified from [12]).

Discussion

Positive *vs.* negative feedback mechanisms

Figure 1 shows that the non-linear increase in signal amplitude from the Early Pliocene to the Late Pleistocene is observed in all $\delta^{18}O$ and SST astronomical components and is related to the long-term mean climate state (global greenhouse and global icehouse). Because the Plio-Pleistocene orbital forcings do not exhibit a similar increase in amplitude, it is suggested that the process controlling the non-linear behaviour of the signal response originates inside the climate system *per se*. The net balance of positive and negative feedback mechanisms is the most likely process behind most of the non-linearities in climate [12-22]. These lagged processes, which are linked to the mean climate state, are also strongly dependent on forcing duration (cycle periodicities) and the involved surface/volume. Indeed, the exponential growth of the orbital variance towards the Late Pleistocene exhibits growth rates related to the orbital period, increasing from half-precession to short eccentricity (Figure 1 in [12]). The asymmetry of the R_s suggests an amplified orbital response, to 400% $\delta^{18}O$, during periods characterised as being a global icehouse state (Mid-Late Pleistocene, with the exception of obliquity after 530kyr), and a damped response, to −100%, during periods characterised as being a global greenhouse state (Pliocene). These results may indicate a net prevalence of positive feedback mechanisms during periods of a global icy-state, with strong amplification effects, and a net incidence of negative feedback processes during periods of natural greenhouse-state, with moderate mitigation effects.

Explaining which prevailing feedback mechanisms may have produced these observational results is a difficult task. Despite significant advances in modelling feedbacks between most components of the climate system, the question of the overall role of feedback mechanisms in the climate system remains unresolved. This is partly due to the lack of a rigorous accounting for these feedbacks in numerical studies and also because both feedbacks can be positive or negative, with great uncertainty about their quantitative impact on different time scales [3,19,23]. During glacial inception, forests were reduced in their areal extent, leading to an increase in atmospheric CO_2 concentration, while during deglaciation, CO_2 concentration should have decreased due to the expansion of forests [23]. Palynological studies attest to major changes in vegetation cover both at glacial and interglacial time-scales [24-26]. During a glacial climate, tropical rainforests are reduced in their extent to tropical seasonal forests in tropical lowlands and to xerophytic taxa in tropical highlands [25]. Boreal forests move equatorward, with a compression and fragmentation of forest zones, while cold steppe vegetation expands [24,26]. These vegetation changes modify rates of carbon uptake, which leads to damping of the radiative forcing, both during glacial and interglacial stages, but also terrestrial surface albedo, which acts in the opposite direction on climate [2,3,23]. Albedo is an important

positive feedback mechanism that is able to amplify the climate response. It consists of several components (terrestrial ice-snow albedo, sea ice-snow albedo, vegetation albedo, subtropical desert albedo) which are difficult to quantify [19,23]. The oceans are an important sink for CO_2 through absorption of the gas into the surface water, but the ability of the ocean to remove CO_2 from the atmosphere decreases with increasing water temperature (solubility pump) [7,27]. The efficiency of this positive feedback is dependent on ocean circulation. Warm ocean slow-down the ocean circulation, which reduce the oceanic uptake of CO_2, but increase the storage of carbon in the ocean. However, the net balance of the solubility pump and ocean circulation appears to be positive [28]. Negative feedback mechanism is the average increase in precipitation rate at medium-high latitudes in a greenhouse climate. Such an increase may intensify rates of weathering as fresh water is flushed more rapidly through soil and sedimentary rocks [29]. Removal of atmospheric CO_2 is accomplished through burial of organic carbon and increasing river discharge may lead to a more efficient marine burial of organic carbon [29]. The oceanic carbon pump exerted by primary productivity can act as possible damping mechanism, but this process requires a better understanding. The biological pump consumes CO_2 and transports products of photosynthesis to the deep oceans [7]. Indeed, on orbital time-scales, the depositional conditions of the Plio-Pleistocene Mediterranean sapropels exhibit warm and wet precession-related changes of phytoplankton productivity by river runoff increasing and bottom water anoxia [4,5,30], suggesting a negative feedback on the carbon cycle. In contrast, Shoenfelt et al. [6] strengthen the notion of an uptake of oceanic CO_2 by phytoplankton fertilization during glacial and more dry periods, and their results indicate that glacial physical weathering increased the proportion of highly bioavailable $Fe^{(II)}$ in dust that reaches the sub-Antarctic Southern Ocean, which implies a positive feedback process.

Wide forest cover during the Zanclean and Piacenzian

After the Zanclean (5.33-3.60Ma) and towards the Late Pleistocene, a progressive impoverishment of forests occurred, with an increase in herbaceous taxa following the long-term cooling and drying climate trend [24,26,31,32]. Indeed, the Zanclean, and to a slightly lesser extent Piacenzian (3.60-2.58Ma), were periods characterised by high atmospheric greenhouse gas concentrations and high mean global temperature/rainfall [26,33-35]. During the Zanclean, northern and central Italy were covered by vegetation dominated by hygrophilous and thermophilous forest taxa typical of a humid subtropical-to-warm-temperate climate. Precipitation was sufficiently high for the persistence of a "broad-leaved evergreen/warm temperate mixed forest" to 3.5Ma [26]. A pollen-based biome reconstruction in the Far East Russian Arctic illustrates the changes that transformed a predominantly forested Pliocene Siberian Arctic ecosystem into a largely forest-free environment [31]. In northern Iceland, the Tjörnes section was deposited in a shallow marine environment before 4Ma

[32]. Diverse plant communities, with lowland deciduous forests and hinterland Gymnosperm forests, persisted throughout the Zanclean in a maritime temperate climate with warm summers, comparable to the present west European climate. Rees-Owen et al. [36] reconstructed past temperature and vegetation during the East Antarctic Ice Sheet retreat (mid-to-late Neogene, 17-2.5Ma) using chemotaxonomic study of terrestrial sediments of the Transantarctic Mountains. They concluded that the mean continental summer temperature was ~5°C, warm enough to have allowed woody vegetation to survive and reproduce, including during the austral winter. Biomarkers from vascular plants indicate a low diversity flora consisting of higher plants, moss and algal mats growing in microenvironments in a glacial outwash system. Abietane-type compounds were abundant in some samples, indicating that conifers, most likely Podocarpaceae, grew on the Antarctic continent well into the Neogene. Salzmann et al. [35] provide an overview of global Piacenzian vegetation environments, whose main points are summarised in the following lines:

a) Tropical and sub-tropical vegetation zones: Palaeobotanical data available for the tropical zones of Africa, Australasia, and South America suggest that the evergreen rainforest belt was nearly in the same latitudinal position as today. The adjacent tropical woodlands, savannas, and shrublands expanded and covered much larger areas than today. The higher precipitation caused a retraction of subtropical deserts and semi-desert. Most of the modern arid and semi-arid climate zones in Africa, central Australia, and the Arabian Peninsula were covered during the Pliocene with temperate and tropical xerophytic shrublands and grasslands.

b) Temperate and warm-temperate zones: Warm-temperate evergreen coniferous forests covered the western and southernmost coastal zones of North America, indicating warmer or near modern temperatures. Warm–temperate broadleaved forest became the dominant vegetation in most parts of Europe and Southeast Asia throughout the Pliocene. The modern boreal conifer-dominated forest zone of central and northern Europe was covered during the Pliocene with cool-temperate mixed forests, implying significantly higher mean annual winter and summer temperatures than today. The Piacenzian temperate forests extended in northeastern America and Norway to 60°N and 70°N, respectively.

c) Polar and sub-polar vegetation zones: In northeast Greenland, elements of taiga forest reached 82°N during the Piacenzian. Vegetation reconstruction of the Northern Hemisphere high latitudes indicates that the polar tundra was replaced in most areas by taiga forests with dominant *Picea* and *Pinus*.

Conclusion

The Pliocene wide-expansion of forest cover towards high latitudes, along with a possible high ocean primary productivity and organic carbon burial, could increase the quantitative role of the CO_2-related negative feedbacks to the positive mechanisms like vegetation-snow albedo and solubility pump, causing a gradual, significant, mitigation of the orbitally-paced climate responses (to -100% global $\delta^{18}O$ and equatorial SST R_s). In the Mid-Late Pleistocene, an Earth's icy-state led to a significant reduction in global forest cover, making the negative feedbacks quantitatively less important while emphasizing the sea-ice and ice-sheet albedo feedback in synergy with the vegetation-snow albedo feedback, which tend to amplify global mean temperature changes.

Declaration of Funding

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