Elevation-dependent temperature response in early Eocene using paleoclimate model experiment

Pratik Kad $^1$, Manuel Tobias Blau$^{1,2}$, Kyung-Ja Ha$^{1,2,3,*}$ and Jiang Zhu$^4$

$^1$ Department of Climate System, Pusan National University, Busan, Republic of Korea
$^2$ Center for Climate Physics, Institute for Basic Science, Busan, Republic of Korea
$^3$ BK21 School of Earth and Environmental Systems, Pusan National University, Busan, Republic of Korea
$^4$ Climate and Global Dynamics Laboratory, National Center for Atmospheric Research, Boulder, United States of America

* Author to whom any correspondence should be addressed.
E-mail: kjha@pusan.ac.kr

Keywords: mountains, Eocene, paleoclimate modeling, elevation-dependent warming, future paradox, CO$_2$

Supplementary material for this article is available online

Abstract
Mountains become warmer with elevation in response to greenhouse gas warming, an effect known as elevation-dependent warming. The Eocene is considered a replica of the future climate in an epoch with high atmospheric carbon dioxide concentration (CO$_2$). Therefore, the topographic features of the Eocene strata are of interest. However, obtaining proxy data for mountain regions during the Eocene hothouse is challenging. Paleoclimate model simulation is an effective tool for exploring past climate change. Therefore, we conducted sensitivity experiment simulations employing the Community Earth System Model version 1.2 forced by proxy-estimated CO$_2$ levels. This is the first Eocene study demonstrating the elevation-dependent temperature changes and illustrated using the surface energy budget decomposition. Here five major mountain ranges have been chosen based on their paleogeographic continental location. We found a nonlinear response of elevation-dependent temperature change to CO$_2$ concentrations regulated by seasonal variations. The radiative and non-radiative feedback compensation is responsible for the elevation-dependency temperature changes. Our results suggest temperature perturbations regulate elevation-dependent changes in skin temperature through a combination of feedback under greenhouse warming in the early Eocene. These findings also show future paradox response exhibiting elevation-dependent cooling overall mountain regions due to lower elevation warming.

1. Introduction
Mountains are an essential component of the climate system; they regulate regional and global climates [1]. Mountain ranges are not only important for the hydrosphere and the atmosphere but also for the biosphere. Due to their altitude, a wide variety of ecosystems and biotopes for rare animals and plants can develop on the slopes of mountains which contribute to the regional biodiversity [2–5]. Although montane and alpine regions are the coolest parts of the land surface, evidence suggests that these regions become warmer monotonously under climate change [1, 6–9]. The majority of high mountains present elevation-dependent warming (EDW) pattern with increased warming with elevation [10, 11]. However, EDW is a regional phenomenon that is not yet fully understood in the context of global warming. Many previous studies have presented well-documented evidence for EDW from observations and modeling for both the present and future in various mountainous regions (e.g. [12–19]). Multiple mechanisms [7, 10] can lead to EDW, and a combination of these mechanisms is responsible for the heterogeneous regional patterns of EDW.

Previous studies have discussed EDW in modern times under climate change using observations, reanalysis, or model output, indicating that various physical processes occurred at different temporospatial scales, potentially providing explanations for the effects of topographic features. These processes were influenced by global warming signals [13, 20]. Often,
changes in surface albedo and the related alterations of the net shortwave radiation were listed as the dominant drivers of EDW. However, other processes also contributed to the warming/cooling pattern over complex topography. These include cloud radiative forcing, the effects of changing water vapor contents, the radiative effects of aerosols, and the effects of land cover change [20–22]. In addition to radiative forcing, the heat stored within land surface, ocean surface, or atmospheric circulation can potentially contribute to regional warming [21]. Some studies have pointed out potential mechanisms, such as the latitudinal location of mountain regions or relations to amplified warming in polar regions [12, 21]. On the other hand, some studies suggested an elevation-dependent cooling (EDC), implying more cooling at high elevations compared to low elevations [23, 24]. EDC appeared in response to volcanic eruptions in the last millennium [25]. However, comprehensive studies on the warming climate of past epochs are scarce.

Here, for the first time, we attempted to investigate the sensitivity of the elevation-dependent temperature changes in an early Eocene simulation experiment with varying atmospheric CO$_2$ concentrations. The Eocene was a warm period in the Earth’s history, that occurred approximately ~56–48 million years (Ma) ago. At that time, atmospheric CO$_2$ concentrations exceeded 1500 parts per million by volume [26]. The Eocene era started with the Paleocene–Eocene thermal maximum (~56 Ma ago). The global mean surface temperature raised from 5 °C to 9 °C because of the high atmospheric CO$_2$ concentration [27, 28]. This era is considered a replica of future climates owing to the high projected greenhouse gas concentrations [29–31]. Paleoclimatic model simulations [32–34] have thus been used to understand past climate change and its features.

Global climate models (GCMs) offer a tool to replicate possible climate outcomes in future or past, when observations were not available, in response to changing CO$_2$ concentration. Model sensitivity scenarios forced by data of palaeo climate proxies were found to replicate the past climate reasonable well, with lower temperature gradients in the tropical sea surface temperatures being warmer than those in modern times [32, 35–40]. Using such an approach, we can answer some challenging questions, such as how sensitive elevation-dependency on temperature was to increasing CO$_2$ levels in the past? The main objective of the present work is to explore elevation-dependent temperature features in early Eocene simulations under an extreme warming condition using the Community Earth System Model version 1.2 (CESM1.2) [41] model simulation. This allows us to investigate both elevation dependency of various climatic and thermal conditions, regionally and globally, and its sensitivity to a large varying CO$_2$ concentration in the atmosphere.

2. Methods

2.1. Model experiment

To study the elevation-dependent temperature changes during the Eocene hothouse, we investigated CESM1.2 model simulations which is available at the deep-time model intercomparison project framework (DeepMIP, described in [33]). All participating modeling group experiments were forced by proxy-estimated CO$_2$ levels. The model simulation was performed with CO$_2$ concentrations from 280 ppm (pre-industrial levels) to 2520 ppm (9 × CO$_2$). As our ultimate goal was to investigate elevation-dependency on the temperature at various atmospheric CO$_2$ concentrations, we used CESM1.2 simulations for the following reasons. First, its sensitivity experiments operate with atmospheric CO$_2$ concentrations ranging from 1 to 9 times the pre-industrial value, under paleogeography with vegetation type conditions, and river routing [42] and preindustrial levels of other gases and orbital forcing. Second, this model shows an overall agreement with proxy records and captured realistic greenhouse warming features during the early Eocene [40, 43]. Even though proxy-estimated data during Eocene are limited, the model simulated zonal mean temperature over land illustrates consistency among corresponding proxy estimates (supplementary figure 1). CESM1.2 configuration used in this study consists of the Community Atmosphere Model 5.3, the Community Land Model 4.0, the Parallel Ocean Program 2, the River Transport Model, the Los Alamos sea ice model 4 and a component coupler [41].

All CESM Eocene simulations were conducted on the horizontal resolution of 1.9° × 2.5° over the atmosphere and land with 30 hybrid sigma-pressure model levels in the atmosphere. The ocean and sea ice at a nominal 1° grid with 60 vertical model levels in the ocean. All CESM model simulations have been integrated for 2000 model years, except for 1 × CO$_2$, which ran for 2600 model years. To make the model suitable for a paleoclimate simulation under a high CO$_2$ scenario, the model radiation code has been slightly upgraded to represent missing diffusivity angle specifications [40, 44].

In addition to the different CO$_2$ experiments (1 × CO$_2$, 3 × CO$_2$, 6 × CO$_2$, and 9 × CO$_2$), a pre-industrial simulation based on the CMIP5 standard was considered for comparison. The pre-industrial simulation 1 × CO$_2$ attempted to represent the preindustrial climatic conditions and 4 × CO$_2$ were conducted to represent the future conditions. However, DeepMIP simulations starting at CO$_2$ concentration equal to the pre-industrial conditions assumed a planet without a permanent ice sheet or sea ice cover at the poles [45]. Here, we have considered the climatological mean state from the equilibrated last 100 model years of each simulation.
2.2. Land contrast

2.2.1. Land amplification

The land amplification factor was calculated [46] using the ratio of land warming per unit of zonally averaged warming of the adjacent oceans (Globally, longitudes from 0° to 360°),

\[ A = \frac{\Delta T_L}{\Delta T_o} \] (1)

where \( \Delta T_L \) is the change in local air temperature over masked land (latitude from −90° to 90°), and \( \Delta T_o \) is the zonally averaged change in air temperature over the ocean.

2.2.2. Polar amplification

The polar amplification factor is expressed [47] as the ratio of the change in temperature over land to the corresponding change in global mean temperature,

\[ A = \frac{\Delta T_{gm}}{\Delta T_o} \] (2)

where \( \Delta T_{gm} \) is the change in the global mean temperature.

2.3. Mountain regions and elevation-dependency of the variable

The model elevation topography was available for each grid resolution used in the model simulation. High mountains were assumed to have an orographic elevation of more than 1 km. Therefore, we screened out orographic regions that exceeded 500 m or more in height (figure 1). For the present analysis, we selected the five major high mountain ranges during the Eocene era, as shown in figure 1(b). The regions were the North American mountain range, the Asian mountain range, the Indian Island mountain, the South African mountain, and the Antarctic mountain range. To derive the elevation dependence of warming, we created vertical profiles using equally sized altitude bins with 500 m intervals. To make a bin, we detected the model elevation at each elevation grid point and clustered them starting from 500 m to the mountain summit and taken average at that mid-class interval (supplementary figure 2). Depending on the size of the respective region and the estimated elevation in the paleogeography, the size and number of bins may vary. The mean temperatures of the grid points in each elevation bin were calculated and changes from the sensitivity experiments relative to the preindustrial simulation are represented by lines in supplementary figure S2. The elevation-dependency of the change in the variables was computed by taking the slope between the variable and elevation over each individual mountain range. Least-squares regression was utilized for calculating the slope.

2.4. Surface energy budget decomposition

The surface energy budget analysis used to investigate the causes of elevation-dependent temperature sensitivity can be written in the following form:

\[ Q = S_↓ + S_↑ + F_↓ + F_↑ + (S + L) \] (3)

where \( Q \) denotes land surface heat storage and ground heat fluxes, \( S_↓ \) is the downwelling shortwave radiation, \( F_↓ \) is the downwelling longwave radiation, \( F_↑ \) is the longwave radiation from the surface, and \( (S + L) \) is the turbulent sensible and latent heat fluxes, respectively. Here all fluxes were described as positive (negative) when upward (downward). The equation (3) can be rewritten as:

\[ Q = (1 - \alpha) S_↓ + F_↓ - \epsilon \sigma T_s^4 + (S + L) \] (4)

where \( \alpha \) is the surface albedo considered as a ratio of upwelling to downwelling shortwave radiation for clear-sky conditions. \( F_↑ \) can be expressed as \( \epsilon \sigma T_s^4 \) as a function of the surface radiative/skin temperature \( T_s \), the approximated surface emissivity (\( \epsilon = 1 \)), and the Stefan–Boltzmann constant (\( \sigma \)).

Considering changes in the surface energy budget with respect to changing CO\(_2\) concentrations, following the previous methodology [48, 49],

\[ 4 \epsilon \sigma T_s^3 \Delta T_s = \Delta [(1 - \alpha) S_↓] + \Delta F_↓ - \Delta Q - \Delta (S + L) \] (5)

In equation (5), \( \Delta \) denotes the change in the high CO\(_2\) scenario (3×, 6×, and 9×) with respect to 1× CO\(_2\).

Lu and Cai [48] defined the cloud radiative effect (CRE) as the difference between the all-sky and clear-sky radiations at the surface,

\[ \Delta \text{CRE} = (1 - \bar{\alpha}) \Delta S_{↓, \text{clr}} + \Delta F_{↓, \text{clr}} \] (6)

where overbar denotes 1× CO\(_2\) mean state. \( \Delta S_{↓, \text{clr}} \) and \( \Delta F_{↓, \text{clr}} \) calculated by taking difference between all-sky condition and clear-sky condition radiative fluxes.

Using equation (6), we can rewrite equation (5) and divide by \( 4 \epsilon \sigma T_s^3 \) will give us the surface skin temperature changes into partial temperature contributions, resulting in the linearly decomposed form as follows:

\[ \Delta T_s = \frac{-(\Delta \alpha) (\bar{S}_↓ + \Delta S_↓) + \Delta \text{CRE} + (1 - \bar{\alpha}) \Delta S_{↓, \text{clr}} + \Delta F_{↓, \text{clr}} - \Delta Q - \Delta (S + L)}{4 \epsilon \sigma T_s^3} \] (7)
Each term on the right-hand side of equation (7) contains the partial temperature change due to the corresponding CO$_2$ scenario, which are surface albedo feedback (SAF), the CRE, changes in shortwave radiation for clear-sky not related to SAF (SWCS), longwave radiation for clear-sky (LWCS), land heat storage (HSL), and surface turbulent fluxes (FLUX).

We performed the above calculation at each model grid. Furthermore, we illustrated it in the form of elevation dependency as our study mainly emphasizes elevation-dependent temperature response.
3. Results

3.1. Paleogeography and land warming in the Eocene

A significant difference in the land-sea distribution was evident in the paleogeography [42, 50] compared with a modern topographic map (figure 1). According to the plate tectonics theory [51], continent orientation and ocean bathymetry differ. In the Eocene, mountains were less elevated, and their orographic features were quite different from those in the Anthropocene. Large mountain chains were distributed globally, from the tropics to the poles, and were more concentrated over the mid-latitudes. Figure 1 shows the global mountain ranges in Eocene and pre-industrial simulation. For instance, the highest mountains were located in North America (Rocky Mountains) and the Antarctic during the Eocene, with vegetated land surfaces devoid of a permanent ice sheet [33, 42]. As the Indian subcontinent was not yet connected to Asia, the Tibetan Plateau and Himalayan orogenesis had not occurred. Instead, mountains were distributed on the eastern side of Asia. South Africa had a distinct orography.

The specialty of paleogeography is not just limited to topographic features but also land–sea contracts. Land–sea contrast is a fundamental aspect of the climate system that triggered regional impacts [52, 53]. Figures 2(a) and (b) shows overall land, and polar amplification, respectively, in response to $9 \times CO_2$, the scenario with the highest $CO_2$ concentration in the sensitivity experiment (and more information can be found in supplementary table 1). This indicates that there was more prominent warming over land than compared to the ocean, implying that moistening rate with warming over the ocean was greater than that over land (figure 2(a)). The atmosphere gets moister with the ocean warming as the amplification factor is positive, which is nothing but land warming per unit zonal mean warming of the surrounding oceans. The polar amplification response to $9 \times CO_2$ was also evident with land amplification, with more coverage toward the poles (figure 2(b)). The zonal mean land amplification spread can be seen in the sensitivity experiments. However, the high-elevation warming response (figure 2(e)) was comparatively lower in the context of land amplification, except at the South Pole. This also authenticates that the mean global mountain warming rate is the same in all Paleocene $CO_2$ responses and nearly consistent with future $CO_2$ responses, revealing the importance of elevation topography (supplementary table 1). Therefore, we need to carefully analyze the climate response of mountains on a regional scale. Our primary objective is to explore the elevation-dependent temperature response in the early Eocene and assess its sensitivity to $CO_2$.

3.2. Elevation-dependent temperature sensitivity

The elevation-dependent temperature change was evident in both hemispheres (figure 3). However, its sensitivity differed across regions and, with $CO_2$ concentrations. Here, the slope estimation method indicated the elevation dependency as a linear function, which may be different from that in the real world. The EDW was explicitly amplified at $3 \times CO_2$ over Asia, Africa, and the Antarctic. The EDW exists over Asia, Africa, and the Antarctic under $3 \times CO_2$, furthermore heightened under $9 \times CO_2$ response. In addition, the chaotic shift from EDW to EDC was observed in the Indian plate (figure 3(c)). In contrast, North America shows both flavors of elevation-dependency with less sensitivity under $CO_2$ concentrations; it shows EDC in $3 \times CO_2$, but an EDW signal emerged with increased $CO_2$. Such mixed behavior of elevation-dependent temperature changes among the mountains underscores the complexities across the globe. These changes range from $-0.5 \degree C$ to $4 \degree C$ km$^{-1}$, slightly negative to strongly positive across the six mountain regions. Interestingly, this temperature gradient seemed more tilted towards the Southern Hemisphere’s response to Antarctic amplification. These elevation-dependent changes in surface temperature showed a nonlinear response with more sensitivity to higher $CO_2$ concentrations.

Figure 4 depicts elevation-dependent temperature changes using surface air temperature ($T$) and skin temperature ($T_s$). Both temperature changes have notable seasonal variations with an almost identical pattern in the elevation-dependent temperature response, pointing to a close connection between them. The monthly variation increased with higher $CO_2$ concentrations over each mountain range, and seasonally the reversed response can be noticed at some places. For instance, North America shows EDC during winter, with small values replicated with the $CO_2$ scenario. However, EDW in summer in higher $CO_2$ scenarios counteracts the EDC resulting in an altogether EDW in North America (figure 4(a)). The mountain ranges in Asia and Africa experienced EDW in all seasons with different intensities and sensitivities to increasing $CO_2$ levels, showing very high values and sensitivity in summer, with opposite trends in winter (figures 4(b) and (d)). Similarly, Antarctica showed the most substantial EDW seasonality. In this region, the EDW was the strongest having its highest peaks in the austral summer (December, January, and February), with especially high sensitivity in January. In the austral winter, values are near zero or showed slight EDC (figure 4(e)). In contrast, the Indian plate experienced a decrease in EDW with increased $CO_2$ concentration ($3 \times CO_2$ and $6 \times CO_2$) and EDC ($9 \times CO_2$). These effects were reversed in the scenario with the highest $CO_2$ concentration and weak seasonality (figure 4(c)). This seasonal cycle showed
the dependency of surface air temperature on the skin temperature CO$_2$ forcing. Surface skin temperatures would not respond similarly to surface air temperatures because of the impact of large-scale circulation [46]. The pattern was almost the same for both temperatures; it was more intense in the case of skin temperature. Surface air temperature may be controlled by large-scale circulation; it usually follows the skin temperature, depending on the turbulent fluxes from the surface. These fluxes play an essential role in air–land surface interactions. The air–land surface temperature change and its interaction can be expressed as an imbalance in the surface energy. The skin temperature over land is strongly coupled with sensible heat flux, which is part of the non-radiative feedback.

Using the correlation method, previous studies analyzed the driving mechanisms on the interannual-to-decadal scales [8, 14, 19, 54] However, it cannot be applied to the seasonal scale and is also limited in sense of partial contribution to temperature. A previous study [22] specifically investigated the heat storage term using the surface energy budget for the Tibetan Plateau. This study found increased net radiation at the surface due to a decreased total cloud, and surface snow is favorable for EDW response to 4 × CO$_2$ over the Tibetan Plateau. We attempted to attribute elevation-dependent changes in terms of
3.3. Partial contributions from radiative and non-radiative feedback

Figure 5 illustrates the decomposed partial temperature contribution of different terms affecting the energy and heat budget in mountain regions, which can help elucidate the main contributors to the changes in elevation-dependent temperature changes. Such approaches have been extensively used to attribute polar amplification [48, 49]. The terms in equation (7) indicate the partial contributions to CO$_2$-induced warming and consist of radiative and non-radiative feedback processes. Among these terms, SAF, CRE, SWCS and LWCS represent the radiative processes and HSL and FLUX indicate non-radiative processes. Asia, Africa, and Antarctica have EDW signals and a decrease in EDW or even EDC on the Indian plate, while North America owns mixed signals with increasing CO$_2$. However, major contributors differed between mountain ranges due to their geographic locations. The LWCS mainly consists of water vapor and cloud feedback [11, 55, 56]. Meanwhile, the SAF is an important contributor to warming in mid- and high latitudes, although it is negligible in low-latitude mountain ranges. Other contributors counteracted this warming trend, including the CRE, SWCS, and LWCS. However, LWCS can seasonally contribute to enhanced EDW. It implies that the topographic process might be enhanced with CO$_2$ forcing and is consistent with earlier studies [22, 57], which also attributed a similar term [11, 58–60] as LWCS. The CRE was dominant in low-latitude regions, including the Indian plate and Africa (figures 5(c) and (d)). The mountains in South Africa experienced warming mainly because of the LWCS which experienced an increase with increasing CO$_2$ concentration. The EDW followed this trend, while other radiative terms did not show high sensitivity (figure 5(d)). In contrast, in North America and Asia, the SAF contributed to EDW, and changes in LWCS with increasing CO$_2$ concentration enhanced this trend. The SWCS imposed EDC. Furthermore, LWCS and CRE did not present clear signs thus contributing to the seasonality of EDW (figures 5(a) and (b)). Over the Antarctic, which has reduced seasonal ice and snow cover in the sensitivity scenarios (3 × CO$_2$ and 6 × CO$_2$), the strongest EDW was observed with high seasonality. The SAF followed the annual amplitude and season of EDW which is related to snow cover changes. The strong EDW response interpreted in the 9 × CO$_2$ scenario might be due to the polar region’s ice/snow cover reduction. A recent study [8, 61] also indicated a current decrease in snow depth over Tibetan Plateau corresponding to EDW. In addition, the LWCS contributed to the warming trend with increasing values with increasing CO$_2$ concentrations (figure 5(e)).

The radiative feedback was strong under CO$_2$ forcing and could explain the EDW pattern and its sensitivity over most of the investigated regions. The results so far describe the radiative feedback effects. Notably, the non-radiative feedback facilitated the
radiative contributions and enhanced them. Their negative values (fluxes from land to atmosphere) indicated surface cooling and vice-versa, which ultimately raises the surface air temperature. For instance, the HSL exhibited a similar trend as EDW in North America. The heat stored in the land mass caused the surface temperature to rise, promoting energy release in the form of FLUX (figure 5(a)). The opposite trend can be seen over Africa, where HSL was negative, promoting energy shortage on the surface and triggering positive FLUX values (figure 5(d)). The increased non-radiative feedback (HSL and FLUX), followed by radiative LWCS over the Indian plate under 3 × CO₂ and 6 × CO₂ scenarios, resulted in EWD from non-radiative processes. However, this could not counterbalance the negative effect of CRE with positive non-radiative feedback, resulting EWC in the 9 × CO₂ scenario (figure 5(c)).

The partial contribution of the different terms followed a distinct seasonality similar to that of EDW, which changes with rising CO₂ concentration (figures 4 and 5). The total elevation-dependent response and most partial contribution terms to temperature were the strongest in the respective hemisphere’s seasonal cycle. Elevation-dependent temperature responses are diverse in each mountains regions with complex local feedback interactions. The SWCS alone did not show this distinct seasonality in any regions and had overall low values. In the summer months, the contribution terms and EDW values were stronger than those in the winter. However, the regions in the Northern Hemisphere did not show seasonality as distinct as those in the Southern Hemisphere. During the summer months in Asia and over the Antarctic CRE and SAF dominated over the radiative terms, respectively, and supported by LWCS, enhanced the seasonal EDW change. The non-radiative terms seemed to counteract these effects (figures 5(b) and (e)). Africa was an exception; here, most terms showed positive values and contributed to
Figure 5. Monthly evaluation of the contribution of partial surface temperature terms to the regional elevation-dependent skin temperature \( T_s \) change during Eocene over (a) North America, (b) Asia, (c) the Indian plate, (d) Africa, and (e) the Antarctic in response to \( \text{CO}_2 \) concentrations. The term at left axis indicates the elevation-dependent skin temperature change, surface albedo feedback (SAF), the cloud radiative effect (CRE), changes in shortwave radiation for clear-sky conditions not related to SAF (SWCS), longwave radiation for clear-sky conditions (LWCS), land heat storage (HSL), and surface turbulent fluxes (FLUX).

warming with elevation. Only slightly negative values were observed for the HSL. The dominant driver for seasonal EDW was LWCS, which was elongated and enhanced by positive CRE (figure 5(d)).

Studies on EDW mainly based on in-situ observation or satellite remote sensing. This is explained by the resolution dependence of micro-physical processes, where resolution dependence of EDW in CESM as thoroughly examined in a previous study using another climate model [19]. Modeling studies supported the assumption, that SAF was relevant in topographic regions where snowline changes [10, 62, 63]. In our paleo-climate setup strongest SAF was indicated for the Antarctic where a melt-down of the entire permanent ice-sheet in the 9 \( \times \) \( \text{CO}_2 \) scenario would cause an EDW of more than 10 K km\(^{-1}\) in the local winter months. CRE were also handled as major contributors, where temperatures are sufficient high [19]. CRE was most important for the EDC on Indian plate adding about \(-6\) K km\(^{-1}\) cooling trend in the 6 \( \times \) \( \text{CO}_2 \) and 9 \( \times \) \( \text{CO}_2 \) scenario, which was not balanced entirely by other processes.

Using present-day-based simulation, all mountain regions (figure 6) reveal EDC in future response to \( \text{CO}_2 \), which is a future paradox. Here chosen regions are slightly different from the Eocene as geography, and continental configuration is distinct in the present day. As the Indian plate is now part of Asia, we consider the Andes mountain range from South America, shown in figure 6(c). These surprising results suggest that future 4 \( \times \) \( \text{CO}_2 \) responses are in contrast to our Eocene framework and the present agreement, which strongly suggests raising \( \text{CO}_2 \) is responsible for EDW in the future [22, 58, 64]. Even though it is a paradox between the past and the future responses, future EDC is dominantly controlled by
LWCS overall mountain regions, underscoring the clear evidence for elevation-dependent water vapor and cloud feedback under future CO$_2$ warming. Numerous studies (e.g. [58, 64]) suggested that LWCS is a potential contributor to EDW; we accomplished this relationship in the past climate during Eocene. Figure 6 shows typical seasonal amplification in radiative feedback and non-radiative feedback. SWCS is the second most dominant feedback after LWCS over North America, Asia, Andes, and Africa. Increasing CO$_2$ in the future can lead to more significant warming at low elevations as we have more elevated mountains compared to the Eocene, resulting in EDC. Many studies also underlined that warming is happening to a certain elevation, supporting our argument. For example, a recent study [20] also infers that the Tibetan Plateau below 5 km exhibit EDW; it did not show any EDW above 5 km or in future climate projections. A study [65] demonstrated Antarctic exhibits low elevation warming in response to 2 × CO$_2$ forcing. Additionally, this past-future linkage concludes that elevation-dependent temperature changes strongly depend on the geographical orientation of mountains rather than CO$_2$ concentration.

### 4. Summary and conclusions

This present work attempted to explore the past climate response over mountain regions worldwide using a surface energy budget perspective. An analysis based on CESM1.2 showed evident land amplification. In addition, all-mountain ranges experienced elevation-dependent temperature changes during the Eocene in response to CO$_2$ perturbation. Elevation-dependent temperature responses are diverse in each mountain region, with complex local feedback interactions. A nonlinear elevation-dependent response could be observed for selected mountain regions with CO$_2$ concentrations, whereas Asia, Africa, and Antarctica have strong EDW among the mountain ranges. However, North America and Indian plates exhibit mixed signals of EDW and EDC with lower values. We demonstrated that elevation-dependent temperature response was sensitive to seasonal variations. Therefore, this seasonality contributed to the elevation-dependency on temperature and its mean strength.

Here, we use the partial temperature decomposition method to understand physical processes that contribute to elevation-dependent changes. Partial temperature decomposition also suggested that the nonlinear effects resulting from a combination [10] of feedback processes were responsible for the elevation-dependent temperature change over the mountains. The study [22] demonstrated that shortwave radiation is the main contributor to EDW in response to 4 × CO$_2$. However, the present study suggests these interactions are complicated because of their locality, as various terms work differently in elevation-dependent processes. An increase in radiative feedback primarily controlled elevation-dependent temperature change, with the Antarctic
exhibiting high sensitivity to CO$_2$. The SAF and LWCS were dominant factors in Antarctic EDW, supported by non-radiative feedback processes. At the same time, the EDC response to $9 \times CO_2$ over the Indian plate occurred despite the negative CRE and SAF supported by HSL. In short, non-radiative feedback supports the radiative adjustment to EDW by enhancing radiative feedback in the early Eocene, and negative non-radiative feedback supported the radiative adjustments to EDW by enhancing radiative feedback. Conversely, a reduction in radiative feedback with positive non-radiative feedback pushes EDC. The equilibrium between radiative and non-radiative processes was responsible for the elevation-dependent temperature changes. We also found a paradox response in present-day based simulation with higher atmospheric CO$_2$ due to the dominant control of LWCS at the lower elevation, resulting in EDC. Previous literature has focused extensively on the EDW in the Tibetan Plateau (the third pole of the earth) in the present and future climate simulations [8–10, 13, 20, 22, 24, 54, 64, 66–68]. However, we found EDC over Tibetan Plateau in response to future CO$_2$, which highlights that elevation-dependent changes strongly depend on the geography of mountains in addition to LWCS. This work also suggests exploring various mountain regions using an energy budget framework can be possible; this can be used to study related elevation-dependent climate change for the past, present, and future.

The regional climate models (RCMs) captures orographic feature better than coarse-resolution GCMs [19, 22, 63, 69]. At the same time, the RCMs has constraints in simulating large-scale features such as global warming. Still, those coarse-resolution GCMs would be valuable in exploring the impact of large scale factors and global warming [11, 19, 22, 64]. The paleoclimate GCMs is usually conducted at a coarser resolution because proxy estimates are limited in numbers and spatial coverage, which also saves computation time. The modeling approach still has some limitations and uncertainty, such as resolution, representation of dynamic vegetation, constant for astronomical parameters, and solar constant. Still, paleoclimate GCMs play a crucial role in understanding paleoclimate features and changes.

Besides, one can resolve the role of polar amplification in global elevation dependency changes for seasonal to interannual scales, which is still an open question for the elevation-dependent climate change research community.

Data availability statement

The Eocene simulations used in this study are available by following the instructions at DeepMIP (www.deepmip.org/data-eocene/).

Acknowledgments

The authors would like to express their gratitude to the two anonymous reviewers for their constructive comments and suggestions. This work was supported by the National Research Foundation of Korea (NRF) grant funded by the Korean government (MSIT) (Grant No. 2020R1A2C2006860). The CESM project is supported primarily by the National Science Foundation (NSF). This material is based upon work supported by the National Center for Atmospheric Research, which is a major facility sponsored by the NSF under Cooperative Agreement No. 1852977. Computing and data storage resources, including the Cheyenne supercomputer (10.5065/D6RX99HX), were provided by the Computational and Information Systems Laboratory (CISL) at NCAR. The authors wish to acknowledge use of the PyFerret program for analysis and graphics in this paper. PyFerret (https://ferret.pmel.noaa.gov/ferret/) is a product of NOAA’s Pacific Marine Environmental Laboratory.

Author contributions statement

P K and K J H conceived this study. P K came up with the initial idea, conducted the investigation, produced figures, and contributed to the initial draft preparation. M B performed data extraction, assisted in the investigation, produced supplementary figures, and contributed to the initial draft preparation. K J H supervised this study, assisted in the investigation, and contributed to writing the draft. J Z performed the model simulations and contributed to writing the draft. All authors interpreted the results and the improvement of the manuscript.

ORCID iDs

Pratik Kad  @ https://orcid.org/0000-0002-0010-0891
Manuel Tobias Blau  @ https://orcid.org/0000-0003-4649-6040
Kyung-Ja Ha  @ https://orcid.org/0000-0003-1753-9304
Jiang Zhu  @ https://orcid.org/0000-0002-0908-5130

References

[1] Hock R et al 2019 High mountain areas IPCC Special Report on the Ocean and Cryosphere in a Changing Climate (Cambridge: Cambridge University Press) pp 131–202
[2] Viviroli D, Dürr H H, Messerli B, Meybeck M and Weingartner R 2007 Mountains of the world, water towers for humanity: typology, mapping, and global significance Water Resour. Res. 43 W07447
[3] Peters M K et al 2019 Climate–land-use interactions shape tropical mountain biodiversity and ecosystem functions Nature 568 88–92
[4] Antonelli A et al 2018 Geological and climatic influences on mountain biodiversity Nat. Geosci. 11 718–25
Byrne M P and O’Gorman P A 2013 Land-ocean warming contrast over a wide range of climates: convective quasi-equilibrium theory and idealized simulations J. Clim. 26 4000–16

Masson-Delmotte V et al 2006 Past and future polar amplification of climate change: climate model intercomparisons and ice-core constraints Clim. Dyn. 26 513–29

Lu J and Cai M 2009 Seasonality of polar surface warming amplification in climate simulations Geophys. Res. Lett. 36 L16704

Boeke R C and Taylor P C 2018 Seasonal energy exchange in sea ice retreat regions contributes to differences in projected Arctic warming Nat. Commun. 9 1–14

Sewall J O, Sloan L C, Huber M and Wing S 2000 Climate sensitivity to changes in land surface characteristics Glob. Planet. Change 26 445–65

Lenardic A, Jellinek A M, Foley B, O’Neill C and Moore W B 2016 Climate-tectonic coupling: variations in the mean, variations about the mean, and variations in mode J. Geophys. Res. Planets 121 1831–64

Kamae Y, Watanabe M, Kimoto M and Shiogama H 2014 Summertime land–sea thermal contrast and atmospheric circulation over East Asia in a warming climate—part II: importance of CO₂-induced continental warming Clim. Dyn. 43 1973–91

Joshi M and Gregory J 2008 Dependence of the land-sea contrast in surface climate response on the nature of the forcing Geophys. Res. Lett. 35 L24802

Guo D, Sun J, Yang K, Pepin N and Xu Y 2019 Revisiting recent elevation-dependent warming on the Tibetan Plateau using satellite-based data sets J. Geophys. Res. 124 8511–21

Cappi P, Brient F, Zelinka M D and Hartmann D L 2017 Cloud feedback mechanisms and their representation in global climate models Wiley Interdiscip. Rev. Clim. Change 8 e465

McKim B A, Jevenanjee N and Vallis G K 2021 Joint dependence of longwave feedback on surface temperature and relative humidity Geophys. Res. Lett. 48 e2021GL094074

Huang Y, Xia Y and Tan X 2017 On the pattern of CO₂ radiative forcing and poleward energy transport J. Geophys. Res. 122 10–578

Rangwala I, Miller J R and Xu M 2009 Warming in the Tibetan Plateau: possible influences of the changes in surface water vapor Geophys. Res. Lett. 36 L06703

Ruckstuhl C, Philipona R, Morland J and Ohmura A 2007 Observed relationship between surface specific humidity, integrated water vapor, and longwave downward radiation at different altitudes J. Geophys. Res. 112 D03302

Naud C M, Miller J R and Landry C 2012 Using satellites to investigate the sensitivity of longwave downward radiation to water vapor at high elevations J. Geophys. Res. 117 D05101

Shen L, Zhang Y, Ullah S, Pepin N and Ma Q 2021 Changes in snow depth under elevation-dependent warming over the Tibetan Plateau Atmos. Sci. Lett. 22 e1041

Asaoka Y and Kominami Y 2012 Spatial snowfall distribution in mountainous areas estimated with a snow model and satellite remote sensing Hydrol. Res. Lett. 6 1–6

Giorgi F, Hurrell J W, Marinucci M R and Beniston M 1997 Elevation dependency of the surface climate change signal: a model study J. Clim. 10 288–96

Palazzi E, Filippi L and von Hardenberg J 2017 Insights into elevation-dependent warming in the Tibetan Plateau-Himalayas from CMIP5 model simulations Clim. Dyn. 48 3991–4008

Singh H A and Polvani L M 2020 Low Antarctic continental climate sensitivity due to high ice sheet orography NPIJ Clim. Atmos. Sci. 3 1–10

Zhu L and Fan G 2022 Assessment and projection of elevation-dependent warming over the Tibetan Plateau by CMIP6 models Theor. Appl. Climatol. 147 1713–23

You Q, Chen D, Wu F, Pepin N, Cai Z, Ahrens B, Jiang Z, Wu Z, Kang S and AghaKouchak A 2020 Elevation dependent warming over the Tibetan Plateau: patterns, mechanisms and perspectives Earth Sci. Res. 210 103349

Guo D, Yu E and Wang H 2016 Will the Tibetan Plateau warming depend on elevation in the future? J. Geophys. Res. 121 3969–78

Meinshausen M, Gutowski W, Jones R, Leung R, Mieghemis S, Nunes A and Qian Y 2009 A regional climate change assessment program for North America EOS 90 311