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Aerosol and Solar Irradiance Effects on Decadal Climate Variability and Predictability

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Abstract The expanding interest in decadal climate variability, predictability, and prediction highlights the importance of understanding the sources and mechanisms of decadal and interdecadal climate fluctuations. The purpose of this paper is to provide a critical review of our current understanding of externally forced decadal climate variability. In particular, proposed mechanisms determining decadal climate responses to variations in solar activity, stratospheric volcanic aerosols, and natural as well as anthropogenic tropospheric aerosols are discussed, both separately and in a unified framework. The review suggests that the excitation of internal modes of interdecadal climate variability, particularly centered in the Pacific and North Atlantic sectors, remains a paradigm to characterize externally forced decadal climate variability and to interpret the associated dynamics. Significant recent advancements are the improved understanding of the critical dependency of volcanically forced decadal climate variability on the relative phase of ongoing internal variability and on additional external perturbations, and the recognition that associated uncertainty may represent a serious obstacle to identifying the climatic consequences even of very strong eruptions. Particularly relevant is also the recent development of hypotheses about potential mechanisms (reemergence and synchronization) underlying solar forced decadal climate variability. Finally, outstanding issues and, hence, major opportunities for progress regarding externally forced decadal climate variability are discussed. Uncertain characterization of forcing and climate histories, imperfect implementation of complex forcings in climate models, limited understanding of the internal component of interdecadal climate variability, and poor quality of its simulation are some of the enduring critical obstacles on which to progress. It is suggested that much further understanding can be gained through identification and investigation of relevant periods of forced decadal climate variability during the preindustrial past millennium. Another upcoming opportunity for progress is the analysis of focused experiments with coupled ocean–atmosphere general circulation models within the umbrella of the next phase of the coupled model intercomparison project.

Keywords Decadal climate variability · Volcanic forcing · Solar cycle · Tropospheric aerosol · Volcanic aerosol · Climate modes · Forced decadal variability · Climate reconstructions · Coupled climate models

Introduction

Decadal climate prediction builds upon the characterization, understanding, attribution, and simulation of decadal climate variability (DCV) as well as on diagnostic and prognostic studies of its predictability. A multi-annual to decadal forecast horizon implies that the complexity and uncertainties arising from the interaction between the climatic response to external forcing and the ongoing internal climate variability must be accounted for. Internal climate variability arises from spontaneous instabilities within the climate system. Its interdecadal component—i.e., variability on time scales roughly ranging from a decade to a century—typically arises from a combination of processes including stochastic atmospheric noise and coupled atmosphere–ocean processes; then, ocean dynamics

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linked to the slow propagation of extra-tropical planetary waves crucially sets the time scale of interdecadal variability in the different ocean basins [1, 2]. The active thermohaline circulation existing in the Atlantic Ocean provides another possible oceanic mechanism of interdecadal climate variability [3]. The relevance of this mechanism is currently subject of debate since a minimal oceanic influence was proposed for the Atlantic Multi-decadal Oscillation or AMO [4–6].

The forced component of DCV is caused by various natural and anthropogenic external drivers whose climatic impacts often cannot be fully understood in terms of simple metrics such as the associated radiative forcing: indirect effects of these forcing agents, as for instance those of anthropogenic aerosols on cloud, snow, and ice properties, are critical to determining their overall climatic impacts [7]. Furthermore, tropospheric and volcanic aerosols feature highly heterogeneous spatio-temporal evolutions, linked to, e.g., seasonal regional patterns of dust, shifting regional patterns of anthropogenic aerosol sources, the irregular and episodic occurrences of wildfires or volcanic eruptions across the globe, and the intertwining of aerosols with the large-scale stratospheric and tropospheric circulation. It is important to distinguish forcing, i.e., instantaneous change in the global radiative budget; rapid adjustments, i.e., indirect modifications of the radiative budget due to fast atmospheric and surface changes that do not depend on the global-mean surface temperature; and feedbacks, i.e., amplification and dampening loops related to changes in climatic variables that operate through changes in the global-mean surface temperature [8, 9]. Our understanding of all three aspects bounds our understanding of forced DCV and its predictability (DCVP): this review therefore also takes into account aspects of the forcing, although it mostly focuses on feedbacks.

Instrumental climate records are crucial for the identification and characterization of the different manifestations of DCV, but are too short and express climate variability largely influenced by anthropogenic activities to robustly infer the full range of natural DCV (for instance, only five strong tropical volcanic eruptions occurred during the observational period). Past climate reconstructions from biogeochemical and documentary data—so-called proxies—and climate simulations with numerical models have thus become central to the characterization and understanding of DCV. On the one hand, inferring a past climate from raw observations of the natural world is a grand challenge: uncertainties enter reconstructions, among other ways, through the dating of the proxy data, through the proxy–climate transfer function, and through the assumption of a relatively stable proxy–climate relationship through time (e.g., [10, 11]). On the other hand, climate models are imperfect and the generality of model results is often affected by the diverse implementations of complex forcings in the various models and the variety of specific experimental designs. Therefore, this review also gives credit to ongoing efforts by the climate research community aimed at improving coordination in comparative assessments between paleoclimate simulations and reconstructions [12] as well as in the design, analysis, and dissemination of relevant climate model experiments' output (e.g., [13–15]).

Modes of climate variability are recurrent regional or hemispheric expressions of internal dynamics within the coupled ocean–atmosphere system with different susceptibilities to external forcings. They are widely accepted paradigms in the study of DCVP to characterize internal variability, to describe forced responses (see, e.g., the recent review about volcanic forcing by Swingedouw et al. [16]), and to assess the skills of decadal forecast systems. This review therefore refers to several prominent climatic modes that are mentioned in the cited papers. It is assumed that the reader is familiar with phenomena such as the AMO, the Atlantic Meridional Overturning Circulation (AMOC), the North Atlantic Oscillation (NAO), the Southern Annular Mode (SAM), the El Niño-Southern Oscillation (ENSO), the Indian Ocean Dipole (IOD), the Pacific Decadal Oscillation (PDO), and the Interdecadal Pacific Oscillation (IPO).

This review privileges papers published since the fifth assessment report of the Intergovernmental Panel on Climate Change (IPCC-AR5). The proposed selection—certainly non-exhaustive—aims at distilling recent advances in the characterization and understanding of DCVP forced by stratospheric volcanic aerosols, tropospheric aerosols, and solar irradiance, considered individually and within a unified framework. It also aims at delineating overarching open questions and possible major opportunities for progress.

Stratospheric Volcanic Aerosols

During a major volcanic event, large amounts of chemically and microphysically active gases and solid particles are ejected into the atmosphere. If the volcanic column penetrates the stratosphere, sulfur-containing gases soon turn into a thin aerosol cloud that distributes in the lower stratosphere, where it typically persists for a couple of years. It is the radiative properties of these enduring aerosol particles that largely determine the climatic relevance of volcanic eruptions: aerosol particles enhance the Earth's albedo through scattering of short-wave radiation, whose direct consequence is surface cooling; they have also a strong greenhouse effect, i.e., they absorb infrared and near-infrared radiation, whose local effect is stratospheric warming. Both radiative effects depend on the characteristics of the volcanic aerosol cloud, such as its aerosol composition, its optical depth, particle size distribution, and height; these in turn depend on the characteristics of the source, including location and season of the eruption, as well as its magnitude [17].

Instrumental observations, coupled climate simulations, and proxy-based reconstructions have corroborated the idea that strong volcanic eruptions, or periods of enhanced volcanic activity, can influence the establishment and evolution of decadal and even longer climate anomalies. Volcanic forcing can impact DCVP in two main ways: first, through the cumulative impact of clusters of small or moderate eruptions on the Earth's radiative balance over decadal (or longer) periods; then, through interannual and decadal dynamic responses induced in the coupled ocean–atmosphere system by individual (or, similarly, clusters of) strong eruptions. In the first case, the impact stems from the irregular and intermittent nature of volcanic forcing linked to the accidental occurrence of volcanic eruptions around the globe. Cumulative volcanic forcing identifies alternating interdecadal periods of reduced (e.g., 1930s–1950s) and enhanced (e.g., 1960s–1990s) volcanic activities [18]. The response is, in this case, largely due to direct shortwave radiative effects: the increased albedo leads to temporary global surface cooling. Accordingly, Santer et al. [19] explain the divergence between simulated and observed temperature evolutions during the early twenty-first century—particularly concerning the so-called hiatus in global warming—as a consequence of the deficient representation of recent volcanic activity in climate models. Further development of this hypothesis invokes volcanically forced intensification of trade winds over the Pacific to explain part of the recent global warming hiatus, and especially rainfall changes concomitantly observed over the western Pacific islands [20]. Zhang [21] suggests that volcanic radiative cooling in the 1960s similarly contributed to shape the interdecadal global cooling phase initiated in the 1940s with the onset of a warm IPO phase.

Slowdown of the global hydrological cycle is expected after major volcanic eruptions, with significant effects particularly in the wet tropical and monsoon regions (e.g., [22]). The hydroclimate response could be cumulative: Winter et al. [18] link the succession of three prominent interdecadal drying phases over MesoAmerica found in a multi-centennial speleothem record to the cumulative effect of three major volcanic clusters in the early nineteenth century, in the late nineteenth–early twentieth century and in the second half of the twentieth century. As this feature is not robustly reproduced by current climate models, it remains an important dynamic aspect to be understood.

Decadal dynamic responses to volcanic eruptions largely stem from decadal-scale feedback loops that encompass large-scale atmospheric circulation, sea ice, the oceanic thermohaline circulation, and associated heat transports. These include positive feedbacks in their abovementioned classic definition: for instance, the "polar amplification" of post-eruption climatic signals—stronger in the Arctic due to its larger exposure to changes in the poleward oceanic and atmospheric heat flows compared to the Antarctic—provides one element of interhemispheric asymmetry to the simulated decadal climate response to volcanic eruptions [23]. For strong tropical eruptions, a top–down pathway of volcanically forced winter atmospheric variability is known since about two decades. Its key element is a tendency for a strengthened stratospheric polar vortex, caused by the enhanced stratospheric aerosol layer in post-eruption winters through a variety of possible mechanisms (e.g., recently: [24, 25]). Then, the downward propagation of the same signal through stratosphere–troposphere coupling has often been reported to project, in the Northern Hemisphere, on a positive NAO pattern (note, the NAO response is not robust in ensemble historical climate simulations since 1850 CE, see, e.g., [26]). Pioneering modeling studies agree on a general framework where such dynamically induced changes in the tropospheric large-scale circulation superpose on the post-eruption radiative surface cooling to trigger anomalous buoyancy fluxes in the polar and subpolar North Atlantic Ocean. The consequent net strengthening of oceanic convection eventually leads to an enhanced AMOC, whose anomalies typically peak 5–10 years after the eruption (e.g., [27–29]). Signatures compatible with such bi-decadal general mechanism have been found in multi-centennial seasonal reconstructions of North Atlantic and European climates [30].

However, whereas this ocean-mediated mechanism provides a simple dynamic framework to understand the North Atlantic pathway behind simulated volcanically forced DCV, the simulated decadal climate response to a certain specific eruption hardly meets with such generalization. A first obvious determinant factor for this specificity is the eruption's strength: the larger the eruption the more robust is the strengthened polar vortex signal [23, 24], a key element to initiate the decadal response cascade outlined above. The eruption's strength can also critically determine the longer term feedbacks: for example, polar amplification of volcanically induced radiative cooling leads to sea ice expansion; if the enhanced insulating effects of sea ice and/or the stabilizing effects on the water column of freshwater export from the Arctic dominate the response in the oceanic convective regions, then post-eruption AMOC weakening occurs (e.g., [31, 32]). The fact that a wide range of AMOC responses to the same volcanic event can be diagnosed in a multi-model ensemble [33] reveals additional possible sources of uncertainty. Among these are model's characteristics and properties of the simulated climate, employed estimate of volcanic forcing, and representativeness of the considered simulations, i.e., how much they reflect the response(s) typically simulated by the model. In particular, the proneness of a climate model to spontaneously generate bi-decadal AMOC variability seems to affect its excitability to the general AMOC response mechanism outlined above [28]. The different time scales of AMOC response in different models (e.g., [27, 29, 31]) further suggest that at least some degree of model specificity stems from the different representations of ocean dynamics, such as differences in seawater mass formation and propagation (e.g., [34]). Ultimately, the severe biases affecting the simulation of North Atlantic decadal variability with current climate models [35] put caveats on the realism of the simulated post-eruption AMOC behaviors.

More subtly, for the same volcanic event, decadal hemispheric temperature responses have similar spread in a multi-model ensemble and in different realizations with the same climate model [15]. The single-model results demonstrate that the decadal climate response to a given eruption is crucially determined by the background climate conditions at the time of the eruption, including presence and strength of additional forcing factors and phase of the ongoing internal climate variability. Background conditions modulate the strength of the feedbacks initiated by the imposed forcing, implying different signals, and different signal-to-noise ratios as well: They are not merely a source of additive noise for post-eruption decadal climate variability but actively influence the mechanisms involved in the post-eruption decadal evolution [36]. Whereas these results concern strong tropical volcanic eruptions, analog decadal oceanic response pathways and dependency of post-eruption DCV on background conditions can be found in simulations of high-latitude events [37–39] and clusters of moderate eruptions [40].

An interdecadal memory linked to the volcanically forced response of the oceanic thermohaline circulation, as suggested by climate models, would also imply that climatic responses are not independent for clusters of decadally paced eruptions [30]. In particular, a recent study suggests that the timing of subsequent volcanic eruptions determines whether they interfere constructively—if they occur around the same phase of internal modes of oceanic variability—or destructively—if they occur around opposite phases of internal modes of oceanic variability [28]. The concept of constructive/destructive interference is schematically illustrated in Fig. 1a.

The decadal response of the extra-tropical North Pacific to volcanic forcing has received less attention than its Atlantic counterpart, arguably also due to lack of robust paleoclimate evidence about past PDO evolution [41]. Wang et al. [42] explain the significant tendency to develop a negative post-eruption PDO phase in the Bergen Climate Model as a response of the extra-tropical tropospheric circulation over the North Pacific Ocean to the same top–down atmospheric mechanism outlined above. However, this behavior does not emerge as predominant in other climate models that describe the PDO as an internal mode resilient to natural forcing [43, 44].

Volcanic eruptions further affect DCVP through their impact on interannual tropical dynamics. Instrumental observations and climate proxy-based reconstructions consistently suggest a tendency for a warm ENSO response in the first post-eruption year (e.g., [45, 46]). ENSO theory entails several key atmospheric and oceanic processes through which changes in the mean climate state can lead to changes in ENSO (e.g., recently: [47]). For instance, the oceanic dynamical thermostat mechanism provides simple theoretical arguments to explain an El Niño response to volcanic forcing: uniform radiative cooling over the tropical Pacific induces weaker equatorial ocean upwelling that partially compensates for the oceanic heat losses; the resulting zonal asymmetry in sea-surface temperatures (SSTs) is then amplified by the Bjerknes feedback, leading to an El Niño. Significantly increased likelihood of an El Niño occurrence after strong volcanic eruptions is found in a recent multi-model assessment that considers major volcanic events of the past 150 years, with response patterns that are overall consistent with the oceanic dynamical thermostat mechanism but with the peak anomaly somehow delayed compared to observations [48]. However, no clear evidence that natural forcing (including volcanic eruptions) affected ENSO variance during the past four centuries is found in recent coral-based reconstructions [49]. Last millennium climate simulations also provide mixed evidence about how ENSO responds to strong volcanic forcing (e.g., [29, 33, 42]). Notwithstanding differences in the models' characteristics and biases, discrepancies between different model results and between simulated and reconstructed evidence could emerge due to a dependency of ENSO's response on background climate conditions [38, 39, 50], sampling issues in simulation ensembles [51] and uncertainty linked to the eruption's season [52].

Maher et al. [48] also report the post-eruption development of a significant zonal SST gradient in the Indian Ocean similar to the Pacific one, which corresponds to increased likelihood of a more positive IOD phase after strong volcanic eruptions. Internal IOD–ENSO dynamics [53] could then be implicated in the increased likelihood found in some models to develop a significant tendency toward La Niña conditions a few years after major eruptions [29, 42].

Overall, assessment of volcanically forced DCV appears to be complicated by many limiting factors, primarily the limited number of observed volcanic events and the substantial uncertainties associated to events in the preinstrumental period, for which current ranges of reconstructed and simulated responses are not sufficiently constrained. Stratospheric circulation changes induced by the volcanic enhancement of the aerosol layer are crucial to initiate the cascade of decadal responses in the couple ocean–atmosphere system, yet forcing mechanisms of the stratospheric response remain poorly constrained. Excitation of internal modes of decadal and interdecadal climate variability remains a paradigm to characterize and understand volcanically forced decadal climate variability: inadequate simulation of such modes can significantly bias our interpretation of volcanically forced dynamics.

Tropospheric Aerosols

In the troposphere, in addition to radiative effects, aerosol particles can indirectly affect climate by serving as cloud condensation nuclei and ice nuclei (e.g., [54]). Estimates of global radiative forcing of tropospheric aerosols remain uncertain

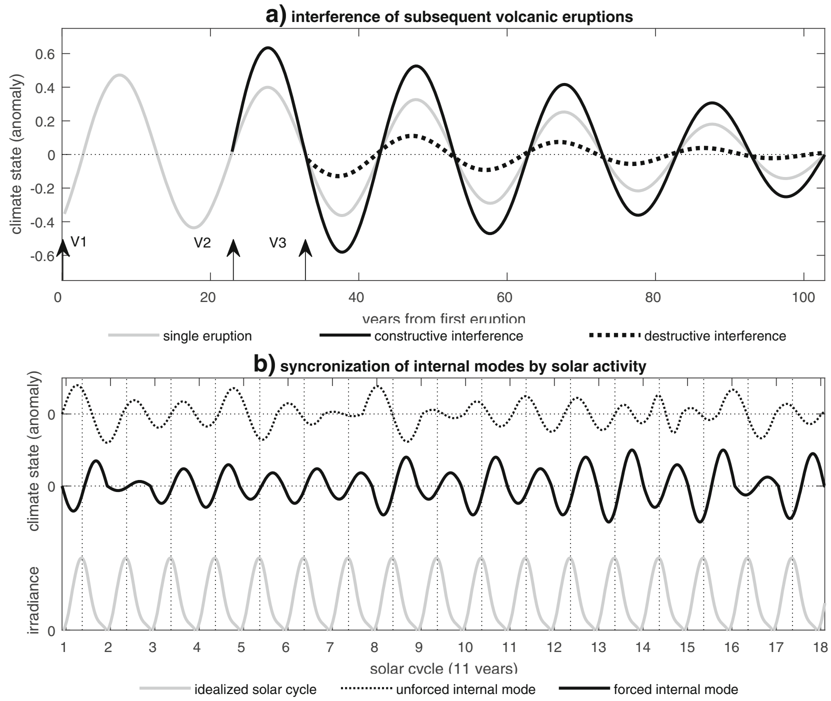

Fig. 1 Schematic representation of volcanic (*top*) and solar (*bottom*) impacts interfering with internal climate variability. **a** Constructive and destructive interference between two subsequent volcanic eruptions on an idealized O (~20 years) AMOC cycle. The eruptions occur at the same AMOC phase for constructive interference (*V1* and *V2*, paced at a 1-cycle interval) and at opposite AMOC phases for destructive interference (*V1* and *V3*, paced at a one and a half-cycle interval). **b** Synchronization of internal variability by solar activity: internal processes spontaneously generate near-decadal variability (*dashed black line*), which tends to get frequency- and phase-locked (with lag, as shown here) in the presence of solar cycles (*continuous black line*)

particularly due to the low confidence on the quantification of the associated cloud adjustment, but they are negative within confidence intervals [7, 8]. Using a simple model for the time history of aerosol forcing, Stevens [55] has recently revisited the lower bound on aerosol forcing, arguing that values less than $-1.0\ \mathrm{Wm}^{-2}$ are very unlikely, hence substantially narrowing the confidence range toward weaker (i.e., less negative) aerosol forcing. Notwithstanding the debate on global aerosol forcing, it is the distributional heterogeneity of tropospheric aerosols that is most important for DCVP. It stems from regional changes in both the source strength and the sink efficiency of natural and anthropogenic aerosols. Aerosol concentrations have peaked between the 1960s and the early 1990s over Europe and North America while they have been increasing in recent decades over Eastern Asia, following the evolution of precursors SO_2 emissions of anthropogenic origin (e.g., [20]). Given the short lifetime of sulfate, this strong spatial and temporal heterogeneity in amounts and properties of tropospheric aerosols implies strong regionalization of the associated radiative forcing (see, e.g., [9]). A key question for historical DCV is, then, how this localized forcing interacted with the global forcing from homogeneously distributed greenhouse gases, given their temporal overlap.

Twentieth century climate simulations suggest that the climatic responses to anthropogenic aerosols and greenhouse gases show both similarities and differences in their spatial pattern: common aspects stem from the similar SST-mediated effects of both forcing types, reflective of the similar ocean–atmosphere feedbacks they trigger [56]; major distinct simulated features of anthropogenic aerosol forcing include enhanced cooling in the Northern Hemisphere's mid-latitudes and southward shift of the intertropical convergence zone [57]. Interpretation of these results should consider that anthropogenic aerosol signals in current climate models are likely distributed much more broadly over the oceans than is observed [55]. Nonetheless, recently unveiled observational evidence substantiates the model-based hypothesis that historical anthropogenic aerosols profoundly affected interhemispheric dynamics via the Hadley circulation [58]. Whether and how anthropogenic aerosol forcing affects historical climate variability simulated in the extra-tropical and polar Southern Hemisphere remain controversial: for instance, the SAM signal forced by anthropogenic aerosols in a multi-model ensemble of single-forcing historical simulations is uncertain and not robust [59]. A possible interhemispheric forcing mechanism by tropospheric aerosols over the Atlantic can be nevertheless envisaged based on the recent discovery that the impacts of the AMO reach as far as the extra-tropical and polar Southern Hemisphere through a Rossby wave bridge [60]. This calls for a more robust quantification of the externally forced portion of observed North Atlantic SST variability in the past century, a matter of unsettled debate in recent years (e.g., [20, 61–63]). Model results also suggest a significant aerosol effect on the multi-decadal evolution of North Pacific SST during the twentieth century, again with some ambiguity about the underlying mechanism (e.g., [64, 65]). In the HadGEM2 model, the forcing pattern is significantly modulated by the large-scale atmospheric circulation associated to internal variability, and the response is largely determined by aerosol–cloud processes [64].

Modeling studies have linked recent regional trends of tropospheric aerosol concentrations to the twenty-first century hiatus although with some diversity concerning relevance and underlying mechanism: Takahashi and Watanabe [20] attribute a (minor) role to tropospheric aerosols as regional modulator of the (dominant) volcanic impacts in the Pacific and Atlantic; Smith et al. [66] invoke the robust correspondence between the regional imprint of anthropogenic aerosols forcing over the Pacific and the negative phase of the PDO, a major trait of the hiatus. Based on climate model experiments, a significant fraction of Arctic warming in the period 1980–2005 can be explained by the concomitant reductions in anthropogenic SO_2 emissions over Europe, with aerosol radiative effects on summer Arctic albedo and meridional heat transport changes as main triggering factors [67]. Accordingly, different assumptions about future aerosol emission estimates significantly affect the simulated evolution of Arctic and global climate through cross-hemispheric impacts on meridional energy transport [68].

Overall, as noted for stratospheric volcanic aerosols, the effects of tropospheric aerosols appear to be tightly intertwined with internal climate dynamics, a connection that becomes even more intimate for natural aerosols (e.g., [69, 70]). On the one hand, robust quantification of observed tropospheric aerosol impacts is obviously complicated by the superposing impacts of the concomitant greenhouse gas forcing. On the other hand, model-based assessments are strongly bounded by the quality of their representation of clouds and aerosol–cloud interactions, and possibly biased by an overrepresented spatial broadening of the forcing.

Solar Irradiance

Solar radiation and fluxes of particles affect both the thermal structure and the chemical composition of the Earth's atmosphere [71]. Variations in solar irradiance occur over a wide range of wavelengths and of time scales, the 11-year solar cycle being the potentially most relevant for DCVP. As decadal variations in the total radiative input are rather small at the top of the atmosphere (roughly 0.1%), amplifying feedbacks are necessary to explain potential solar forced DCV (e.g., [72]). The IPCC-AR5 reports medium confidence that the 11-year cycle of solar variability influences decadal climate fluctuations in some regions of the Earth [73]. Based on this premise, the relevance of the recent advancements in the study of solar forced climate variability discussed below appears even more significant.

An obvious pathway for solar climate forcing of DCVP is through radiatively induced changes in surface temperatures, the so-called bottom–up mechanism: broadening of the Hadley circulation regime with poleward shift of the subtropical tropospheric jets occurs under solar irradiance maxima, largely due to evaporative and convective responses over the subtropical Pacific (e.g., [74, 75]). Climatic signals related to changes in total solar irradiance are not restricted to the tropics, as observational and model-based evidences highlight responses in the extra-tropical North Pacific as well [76, 77]. It must be noted, nonetheless, that the observed evidence for a PDO-solar interaction described in van Loon and Meehl [77] is admittedly based on a small observational sample.

The amplitude of (decadal) irradiance changes is spectrally heterogeneous, and specifically inversely dependent on wavelength (it amounts to up to 10% over an 11-year solar cycle around the ultraviolet 200 nm spectral band). As variations in the photolytic production rate of ozone in the tropical stratosphere are largely due to solar ultraviolet irradiance changes, the associated variations in stratospheric radiative heating rates provide the source for top–down mechanisms of solar impacts on DCV. Increased ozone concentrations under strong solar irradiance warms the tropical upper stratosphere, which leads to stronger meridional temperature gradient between mid- and high-latitudes and associated thermal wind response; stratosphere–troposphere interactions transfer this stratospheric signal to the troposphere, which is observed, in the Northern Hemisphere, as a delayed (by a few years) solar modulation of the NAO variability [78–80]. Atmosphere–ocean coupled processes may be the key for explaining the observed timing and strength of the NAO response to solar forcing (e.g., [80, 81]). Specifically, the concept of reemergence of the solar signal seems to be particularly relevant: each winter, the top–down forcing of the NAO influences also the underlying SSTs; at the end of the winter, the SST anomaly penetrates below the mixed layer, from where it can reemerge in the following winter, thus progressively reinforcing the NAO signal. This NAO mechanism implies a strong regional character of the solar surface imprint, which is particularly evident in the North Atlantic/Arctic and Eurasian sectors (e.g., recently: [78, 82–84]). The top–down mechanism may involve also a tropical pathway, whose fingerprint in the lower tropical troposphere has been, however, recently put into question [85]. Sensitivity experiments for different grand solar minima and model configurations with and without interactive atmospheric chemistry showed that the oceanic response to solar forcing is strongly affected by chemistry–climate interactions in the stratosphere, which amplifies dynamical effects related to ozone and sudden stratospheric warming, leading eventually to predominance of top–down mechanisms on bottom–up mechanisms (i.e., thermal effects) [86].

The anthropogenic depletion of stratospheric ozone that occurred on the second half of the twentieth century has greatly contributed to recent Southern Hemisphere's climatic variations, as its projected recovery is expected to do in the upcoming decades (e.g., recently: [87]). This complicates the diagnosis of top–down solar effects on the SAM in observations and historical simulations. Using a high-pass filter to remove the ozone depletion signal, Gillett and Fyfe [26] found a significant, almost year-round positive SAM response to solar irradiance in a large multi-model ensemble of historical simulations, but with a small signal-to-noise ratio.

As for volcanically forced DCV, the phasing of spontaneous variability of internal modes is essential for the modulation of solar-forced responses in both of Pacific and Atlantic sectors [77]. To this regard, the 11-year solar cycle may act as a synchronizer between spontaneous top–down and bottom–up mechanisms contributing to the NAO variability [88]. Specifically, the proposed synchronization mechanism, identified through climate model experiments, operates through atmosphere–ocean coupled processes (the reemergence) as well as solar modulation of stratospheric dynamics that extends NAO-related variability to the middle and upper stratosphere through deceleration of the Brewer–Dobson circulation and strengthening of the stratospheric polar vortex at solar maxima, and vice versa. Figure 1b schematically illustrates the effect of this synchronization between internal variability and solar activity.

Overall, the emerging picture is that recent research contributed to an improved spatio-temporal characterization of DCV forced by solar irradiance changes and to a stronger grasp on the underlying dynamics. Decadal solar irradiance changes are relatively small and even including the recently proposed reinforcing mechanisms, the amplitude of the forced responses remains generally rather small, despite statistically significant. This could be one explanation why current climate models poorly represent the temporal and spatial signature of the 11-year solar cycle on the surface climate [89, 90]. Nonetheless, a recent statistical assessment of atmospheric blocking over the Euro-Atlantic sector using 1953–2010 winter reanalysis data reveals a significant signature of solar irradiance changes, comparable in amplitude and extent to that of ENSO and the AMO [79]. This result prefigures at least some potential predictability of solar forced DCV, but more research seems necessary to clarify possible interference by other forcing factors.

Outstanding Issues and Opportunities for Progress

As mentioned in the introductory paragraph, understanding of DCVP cannot be separated from characterization and understanding of the forcing, which brings several outstanding issues into light.

Concerning tropospheric aerosols, Stevens [55] suggests that aerosol radiative forcing is less negative and more certain than is commonly believed. Yet, a more complete understanding of aerosol chemistry and physics, hence of aerosol–radiation and aerosol–cloud interactions, is required to better constrain tropospheric aerosol forcing. Moreover, the estimation of precursor natural and anthropogenic emissions is itself subject to uncertainty [69, 91]. Concerning solar activity, the consistency and accuracy of satellite observations of spectral solar irradiance during recent solar cycles remains subject of a strong debate [92, 93]. As a consequence, the estimation and simulation of the 11-year solar cycle signal in stratospheric ozone remains affected by substantial uncertainties [94–96]. Before the satellite era, total and spectral solar irradiances must be reconstructed from proxy-based evidence, a process which carries substantial uncertainty concerning the amplitude and phasing of the 11-year cycles and centennial trends, especially as different empirical irradiance models are in use for the reconstructions [93, 97]. Concerning volcanic eruptions, measurements of stratospheric volcanic aerosols are potentially affected by the limited spatial coverage of satellite observations [98], pushing the necessity of new monitoring strategies particularly near the tropopause. Reconstructed past volcanic activity still suffers by lack of detail and large uncertainties regarding key characteristics of aerosol forcing [15, 99]. Observed and reconstructed solar effects are, in turn, possibly aliased by volcanic signals [85, 100]. In particular, while solar signals are robust in the middle and upper tropical stratosphere, they are not so in the tropical lower stratosphere: there, the portion of observed decadal variability that can be unambiguously linked to the solar cycle may therefore be relatively small compared to volcanic forcing [85].

How to overcome these enduring issues is difficult to envision, but the development of innovative methods for the independent validation of reconstructed forcing estimates is arguably central, as recently exemplified by the assessment of different sunspot number series using cosmogenic isotopic records from meteorites [101]. Also, ongoing coordinated climate modeling activities under the umbrella of the Coupled Model Intercomparison Project phase 6 (CMIP6) [13] not only provide rich opportunities to tackle, in the upcoming years, key questions related to DCVP but also update historical forcing datasets and the associated documentation (e.g., [91, 93]).

Then, a major need is improved implementation of the forcing in climate models. For instance, the representation of the volcanic aerosol cloud remains oversimplified in many of current climate models [17], and the largest uncertainties in the estimates of radiative forcing from CMIP5 historical simulations occur during periods of strong volcanic activity [19]. Similar issues concern tropospheric aerosol forcing: Stevens [55] suggests that the general underestimation of the warming between 1920 and 1950 in current historical simulations reveals the models' tendency to overestimate aerosol radiative forcing, likely due to their overrepresentation of the effects of anthropogenic emissions on clear-sky radiances. Accurate simulation of tropospheric aerosol effects is also hampered by the fact that climate models continue to largely rely on subgrid scale parameterizations to describe cloud microphysics and aerosol–cloud interactions: a great variety of these parameterizations exists, which reflects lack of understanding about some of the relevant microphysical and dynamical processes and of the associated feedbacks with the large-scale circulations [54]. The explicit treatment of aerosol microphysical processes and stratospheric chemistry in climate models can help in improving the simulation of the volcanic aerosol cloud, hence the comparability between observed/reconstructed and simulated responses [102, 103]. Nonetheless, preliminary results from a recent multi-model assessment of volcanic forcing generated by aerosol climate models indicate that associated uncertainties can remain large, questioning which level of model sophistication is necessary to achieve robustness [15].

The obvious solution to surpass all these modeling issues is to develop better and better climate models (e.g., [104]) and experimental strategies. Along this grand long-term goal, comparative assessments between realistic and idealized forcing experiments can help to better evaluate the uncertainties deriving from the imperfect implementation of forcing in current climate models. For instance, proposed mechanisms of solar and volcanic forcing include top–down pathways from the stratosphere to the surface through NAO/SAM changes. This calls for a realistic representation of stratospheric dynamics to simulate the effects of both volcanic (e.g., [25]) and solar forcings [90, 105]. Another milestone not requiring much model development could be a more comprehensive assessment of the simulated sensitivity of the direct radiative and dynamic responses to the spatial structure of volcanic forcing (as shown by, e.g., Toohey et al. [25] and Colose et al. [106]). Concerning solar variability, in particular, it affects the Earth system not only through changes in the radiative energy input but also through energetic particle forcing on the chemical composition of the high-latitude middle atmosphere [107]. Further research is needed to identify the transfer mechanisms of this forcing toward the Earth's surface, and understand its interaction with irradiance changes and with the internal dynamics of the stratosphere. CMIP6 provides, to this regard, a major opportunity for progress, as modeling groups have been encouraged for the first time to address the long-term effects of particles, alongside with effects of solar spectral irradiance variability [93].

Improved understanding of the internal component of DCV and the quality of its simulation is another pillar for robust DCVP research. As illustrated above, aerosol and solar effects on DCVP are intimately connected with the ongoing internal climate variability, and modulation of variability modes in the Earth system is often invoked to explain naturally forced mechanisms of DCV. However, observed characteristics of spatial pattern and time scales of these modes are often not robustly captured in current climate models. This is the case, for instance, for ENSO (e.g., [108]) and the AMO [109], but possibly less so for the PDO/IPO [110, 111]. An obvious opportunity for progress is, in this sense, the detailed assessment of the simulated internal variability of the coupled oceanic and troposphere/stratosphere system in the North Atlantic/Arctic sector (e.g., [112]), given its central role in proposed mechanisms of both volcanically and solar-forced DCVs. Idealized studies of the decadal predictability of the AMO could shed light on its main drivers, and particularly on the role of the AMOC [6]. In the Pacific, improved understanding of the PDO/IPO seems to be the key for robust attribution of recent decadal global temperature changes to external forcing rather than to internal variability [20, 66, 113, 114]. The non-stationary interactions between Atlantic and Pacific variabilities are further necessary subjects of deeper investigation (e.g., [115]). In such a broader perspective, understanding the origins and consequences of systematic climate model biases is a necessary step toward improved overall quality of climate simulations (Eyring et al. 2015). This is clearly the case for decadal climate predictions as well, as recently shown by the fact that the tendency of a decadal climate prediction system to excite a warm ENSO event during the first forecast year affects developing regional systematic errors across the globe [116].

Paleoclimatic evidence is crucial to characterize DCV and infer its general underlying mechanisms, particularly as volcanic and solar effects can be investigated without anthropogenic disturbances. Once again, the similar top–down pathways of solar and volcanically forced DCV require careful study, and possibly a revisited attribution of reconstructed changes during the past several centuries. Understanding of impacts from tropospheric aerosols could as well benefit from a deeper study of preindustrial climates, given the dominant contribution of natural sources to the overall aerosol forcing, and the substantial associated uncertainties [69, 91]. Future DCVP research should bring more focus on key interdecadal periods of the preindustrial era. For instance, the early nineteenth century—a period characterized by the concomitance of two strong tropical eruptions in 1809 and 1815, and of the Dalton minimum in solar activity—is characterized by exceptional interdecadal climate conditions at the global, continental, and regional scales according to many reconstructions (e.g., [12, 49, 117]) and simulations (e.g., [36]). The spatio-temporally heterogeneous and season-specific features described by regional and subregional reconstructions available for this period of the recent past already highlight the difficulties inherent to robust characterization of paleoclimate variability at the level of detail required by the study of DCV. Moreover, the substantial discrepancies existing between simulated and reconstructed features during this period, regarding for instance the Pacific North American pattern [11] and the interdecadal evolution of tropical precipitation [18], highlight how the limitations of currently available paleoclimate tools undermine our understanding of the past DCV. Reconciling simulations and reconstructions for such periods can reveal key information to progress in our understanding of naturally forced DCVP. For this purpose, in light of the strong interconnection between forced and internal components of DCV, a shift in the paradigm for the interpretation of ensemble reconstructions and simulations from a truth-centered—highlighting forced mean responses—to a probabilistic approach—testing how exchanging the reconstruction with any member of the simulation ensemble (or vice versa) changes the probabilistic and climatological characteristics of the ensemble [10, 30]—seems appropriate.

An obvious research goal is the improvement of paleoclimate reconstructions, particularly for the winter season. Millennium-scale transient simulations from climate models provide a long and physically consistent framework where paleoclimate reconstruction methods can be altered and evaluated systematically in absence of the spatial and temporal discontinuities of the real-world climate proxy networks [118]. A few studies have exploited these so-called pseudo-proxy investigations to evaluate available reconstructions of climate modes of variability and explore alternative, more rigorous reconstruction methods (e.g., [11, 62, 119]). Further development and applications of this tool could be decisive to increase fidelity in the reconstruction of complex climate modes, such as the PDO [2]. Critical information can be also acquired from coordinated paleoclimate research activities, such as the 2k Network initiative of "Past Global Changes" [120]: similar efforts with a narrower temporal focus (for instance, covering the past two centuries) could provide a leap forward in terms of spatial and seasonal coverage of the reconstructions. Much needed is also high-resolution and precisely dated information about past marine environments: for instance, crucial progress is promised by the new sclerochronological records that are currently getting acquired in the extra-tropical North Atlantic Ocean [121]. In the North Atlantic, a stronger focus on subpolar gyre dynamics and variability could be instrumental in understanding oceanic responses to strong volcanic forcing beyond the AMOC (e.g., [40]).

A stronger emphasis on the Southern Hemisphere could as well greatly benefit our general understanding of naturally forced DCV. Recent studies tackling diverse aspects of Southern Hemispheric climate variability highlight the importance of better understanding ENSO, SAM and their connection, particularly in the light of their recently observed trends: they appear as a key to explain recent interdecadal variations of Southern Ocean SSTs [122, 123] and regional trends in Antarctic sea ice [124], to better understand recent trend and variability in the Southern Hemisphere's mid-latitude westerlies [125], and to reconcile observed and simulated responses to volcanic forcing [126, 127].

Finally, despite the reported recent advancements in our general understanding of naturally forced DCV, the actual impact of aerosols and solar activity in decadal predictions remains weakly explored. Bellucci et al. [128] discuss the role of initialization of non-oceanic drivers for decadal-scale predictability, and finds some potentials of enhanced initial value predictability for stratosphere and aerosols. Timmreck et al. [129] recently demonstrated that volcanic forcing can significantly affect the prediction skills at the global and regional scales over a 5-year post-eruption period, but the sources of the skills remain unknown. Much can be learned in the future about the implications of volcanic forcing for decadal climate prediction from the experiments proposed by the Decadal Climate Prediction Panel within the CMIP6 umbrella, which includes hindcast experiments with and without major historical eruptions, and 2015 forecast experiments with added volcanic forcing [14]. Studies exploring the impact of a future grand solar minimum could similarly expand our understanding of naturally forced DCVP.

Concluding Remarks

Multiple lines of evidence from climate observations, reconstructions, and simulations point to the potential of volcanic eruptions, tropospheric aerosols, and solar activity to substantially impact on decadal climatic variability. However, uncertainties and gaps of knowledge in the characterization of forced decadal climate responses remain large, and only a few studies have systematically tackled the implication of these forcing agents for decadal predictability and prediction. For all forcing agents, major limitations in understanding arise from incompleteness and shortness of the instrumental observations concerning the forcing as well as the climate response. Further issues concern the deficient representation of key processes in climate models and limitations inherent to reconstructed evidence. Both aspects, nonetheless, provide strong opportunities for progress in the light of increased coordination across observational, paleoclimatological, and climate modeling activities.

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Compliance with Ethical Standards

Conflict of Interest I declare no conflict of interest.

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