1 Introduction

The Atlantic multi-decadal oscillation (AMO), discovered by Schlesinger and Ramankutty (1994) and Enfield et al. (2001), is the dominant mode of Earth’s climate system at the multi-decadal time-scales (Fig. 1). Research have shown that the AMO strongly affects persistent droughts in Africa, such as the famous mega-drought in the 1970s and 1980s, which caused a death toll of more than 100,000 people (Folland et al., 1986; Shanahan et al., 2009; Zhang & Delworth, 2006). It also influences droughts and surface temperature in Asia (Chen et al., 2017; Hao et al., 2016; Li & Bates, 2007; Lu et al., 2006; Wang et al., 2009; Zhang & Delworth, 2006), Europe (Ionita-Scholz et al., 2013; Sutton & Hodson, 2005), North America (Curtis, 2008; Hu & Feng, 2008; Nigam et al., 2011; Sutton & Hodson, 2005), and South America (Chiesi et al., 2009; Kayano & Capistrano, 2014). The AMO significantly modulates Atlantic hurricane activity (Bell & Chelliah, 2006; Caron et al., 2015; Goldenberg et al., 2001; Klotzbach, 2011; Klotzbach & Grey, 2008; Knight et al., 2006; Zhang & Delworth, 2006)Grey, as well as the changing rate of global sea level (Chambers et al., 2012; Frankcombe & Dijkstra, 2009; Jevrejeva et al., 2008; Woodworth et al., 2009). The AMO also affects daily public health such as the asthma mortality rates in the United States (Bonomo et al., 2019).

The AMO index is often defined as the area-averaged SST for the North Atlantic Ocean with linear trend removed (Enfield et al., 2001). An alternative method is to remove

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ATMOSPHERE-OCEAN 60 (3–4) 2022, 307–337 https://doi.org/10.1080/07055900.2022.2086847

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the ensemble mean North Atlantic SST from a large set of climate model historical experiments driven by observed external forcings, which is intended to remove the anthropogenic global warming signals and keep only the natural variability (Knight, 2009; Ting et al., 2009). The two methods often give qualitatively similar AMO indices.

A long-term difficulty in AMO research is the short length of modern instrument data, which cover only two and a half AMO cycles. Therefore, many studies use global climate models to understand the physical mechanisms and global teleconnections of AMO. Fortunately, numerous high-resolution paleoclimate records have been collected in the past century. After a large amount of data had accumulated, paleoclimate field reconstructions were developed (Cook & Jacoby, 1977; Fritts, 1965; Stockton & Meko, 1975), leading to, for example, a series of drought maps for different regions around the world (Cook et al., 2010a, 2010b, p. 2015; Palmer et al., 2015). The recent PAGES 2K consortium is the largest international effort in the history of paleoclimatology to reconstruct the climate history of the common era (Anchukaitis & McKay, 2014; PAGES 2k Consortium, 2017), which has led to the latest continent-scale temperature reconstructions (PAGES 2k Consortium, 2013) and the first generation of paleoclimate reanalysis (Hakim et al., 2016;
Long-term AMO indices have also been reconstructed (Grey et al., 2004; J. Wang et al., 2017b). This paper will review the history of AMO research. AMO dynamics will be discussed in Section 2. AMO teleconnections and impacts will be discussed in Section 3. The effect of AMO on global warming hiatus will be discussed in Section 4. A summary and future directions will be presented in Section 5.

2 AMO dynamics

It is well-established that the AMO is associated with the oscillation of Atlantic meridional overturning circulation (AMOC; e.g. Delworth & Mann, 2000; Frankcombe et al., 2008; Knight et al., 2005; Latif et al., 2004), which is part of the global deep ocean circulation – the thermohaline circulation. Ocean currents in the upper 100 m of the ocean are mainly driven by atmospheric surface winds, while ocean currents in the deep ocean are mainly driven by differences in the water’s density, which in turn, is controlled by temperature (thermo) and salinity (haline). The water’s density increases with higher salinity but decreases with warmer temperature. The global ocean is heated at the top by the incoming sunlight and also getting fresh water input at the top from precipitation and continental outflow. Thus the global ocean temperature generally cools down poleward and downward, while the global ocean salinity has a more complicated distribution. The resultant global ocean density increases downward and poleward, making the global ocean a stable system. However, in the polar regions, ocean water gets cold, leading to the formation of sea ice, which makes the surrounding water saltier. The colder and saltier water has higher density and starts to sink. This initiates the global deep ocean thermohaline circulation. The thermohaline circulation was discovered by Wust (1935, p. 1935; Fig. 2a) and Clowes and Deacon (1935), and its global structure was summarized by Gordon (1986, 1991; Fig. 2b), Broecker (1987, 1992) and Rahmstorf (2006; Fig. 2c). Note that the global thermohaline circulations in Atlantic are mostly meridional overturning circulation trapped to the ocean’s western boundary (Fig. 2c). See previous reviews for the observations and theories of the thermohaline circulation and its connection with the AMO (Buckley & Marshall, 2016; Kuhlbrodt et al., 2007; Talley, 2013; Wunsch & Ferrari, 2004; Zhang et al., 2019).

Theoretical studies have proposed two physical mechanisms for driving the thermohaline circulation. The first one is the mixing-driven mechanism (Bjerknes, 1916; Huang, 1999; Jeffreys, 1925; Munk, 1966; Munk & Wunsch, 1998; Oort et al., 1994; Sandstrom, 1908, 1916). The action of winds and tides generates internal waves in the oceans. These waves dissipate into small-scale motion that causes turbulent mixing. This mixing of heat lightens water masses in the deep ocean and causes them to rise in low latitudes. Resulting surface and intermediate waters are then advected poleward into the North Atlantic where they are transformed into dense waters by atmospheric cooling, evaporation, and salt rejection during sea ice growth. These waters sink to depth and spread, setting up the deep water mass of the ocean. Thereby a meridional density gradient between high and low latitudes is established. Ocean models show that the simulated thermohaline circulation is very sensitive to the vertical mixing (Bryan, 1987; Mignot et al., 2006; Simmons et al., 2004; Saenko & Merryfield, 2005; Tanaka et al., 2012; Oka & Niwa, 2013; Melet et al., 2013, 2016; Osafune & Yasuda, 2013; Schmittner et al., 2015; Zhang et al., 1999). Mixing is caused by tides, winds, and waves. Observations have shown that tidal mixing is the leading component of ocean mixing, especially in the deep ocean, and contributes twice as much as diffusive mixing (Egbert & Ray, 2000, 2003; Garrett, 2003; Koch-Larrouy et al., 2010, 2015; Ledwell et al., 2000; Loder & Garrett, 1978; Polzin et al., 1997; Ray & Susanto, 2016; Rudnick et al., 2003; Zaron & Egbert, 2006). Unfortunately, tidal mixing has not been well represented in many climate models, which is currently a hot topic of research (Green & Nycander, 2013; Klymak et al., 2010; MacKinnon et al., 2017). Without tidal mixing, it is estimated that a vertical diffusivity coefficient of $1 \times 10^{-4}$ m$^2$s$^{-1}$ is required to maintain the observed global thermohaline circulation (Munk, 1966), which is much larger than the observed value of $0.1 \times 10^{-4}$ m$^2$s$^{-1}$ (Gregg, 1998; Hibiya et al., 2006). To obtain a more realistic simulation of AMOC, many ocean models overestimate the contribution of diffusive mixing by employing overly large coefficients in the interior ocean (Kuhlbrodt et al., 2007), which likely make the simulated AMOC overly sensitive to surface winds and heat/salt fluxes.

The second mechanism is the wind-driven mechanism (Sloyan & Rintoul, 2001; Toggweiler & Samuels, 1993, 1995, 1998; Webb & Sugino-hara, 2001). They suggested that the oceanic upwelling in the Southern Ocean can be driven by atmospheric winds. The strong westerly circumpolar winds induce a vigorous northward Ekman transport near the ocean surface, and an upwelling associated with the Drake Passage effect, which is similar to coastal upwelling in sub-tropics.

Forced by mixing and surface winds, the thermohaline circulation is amplified by two positive feedbacks: (1) the salinity advection feedback (Stommel, 1961), and (2) the convective feedback (Welander, 1982). As summarized by Kuhlbrodt et al. (2007), the salinity advection feedback is determined by the basin-size density gradients. The AMOC’s North Atlantic Deep Water (NADW) cell transports salt northward, which maintains the waters’ relatively high salinity in the DWF regions in spite of the freshwater input in the northern midlatitudes. Because of the higher density this fosters sinking. Thus there is a positive advective feedback: An ongoing circulation keeps up the deepwater formation, and conversely, with a halted NADW formation, there is no northward salt transport, and hence DWF is suppressed. On the other hand, the convective feedback concerns one specific deep water formation site. There is a net freshwater input at the deep water formation sites in the North Atlantic. As long as convection happens regularly, the
inflowing fresh water becomes denser by mixing with the deep saline waters. This keeps the surface density high and thus fosters convection to occur in the following winter. However, once convection has not occurred for a couple of years, because of, e.g. variability in the surface fluxes, then the fresh water accumulates at the convection site. This decreases the surface density and inhibits further convection events. In this way, convection sites may be switched off.

The positive feedbacks can only enhance the thermohaline circulation in one direction, but cannot cause oscillations such as the AMO. For oscillations to occur, there must be some physical mechanisms generating the switch between acceleration and slowdown of thermohaline circulation. This is the key question for the AMO theories.

The existing AMO theories can be categorized into seven groups: (1) the stochastic forcing theory (Delworth et al.,
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1993; Delworth & Greatbatch, 2000; Griffies & Tziperman, 1995), (2) the North Atlantic Oscillation (NAO) coupled oscillator (Delworth et al., 2017; Latif et al., 2004, 2007; Sun et al., 2015; Sutton et al., 2018; Timmermann et al., 1998), (3) the zonal-meridional oscillator (Dijkstra et al., 2006; Te Raa & Dijkstra, 2002), (4) the delayed advective oscillator (Lee & Wang, 2010; Zhang & Wang, 2013), (5) the ice coupled oscillator (Boning et al., 2016; Dima & Lohmann, 2007; Frankcombe et al., 2010), (6) the cloud-radiation feedback theory (Bellomo et al., 2016; Brown et al., 2016; Cane et al., 2017; Clement et al., 2015, 2016; Yuan et al., 2016), and (7) the aerosol forcing theory (Booth et al., 2012; Knudsen et al., 2014; Yuan et al., 2016; J. Birkel et al., 2018; Wang et al., 2017b). Theories (1)-(5) emphasize internal dynamics associated with THC, theory (6) emphasizes internal heat fluxes associated with ocean mixed layer, while theory (7) emphasizes external forcing.

The stochastic forcing theory considers the AMO as a damped eigenmode of THC excited by stochastic atmospheric forcing (Fig. 4a,b; Delworth et al., 1993; Griffies & Tziperman, 1995). The time-scale of the AMO is determined by the intrinsic time-scale of the eigenmode, which resonantly responds to the corresponding frequency band in the white spectrum of forcing. The restoring force comes from the surface heat transport, which leads the ocean temperature by about a quarter cycle (Griffies & Tziperman, 1995, their Fig. 8).

The NAO coupled oscillator theory considers the interaction between THC and NAO. The THC-related sea surface temperature anomalies drive NAO anomaly, which in turn causes changes in associated winds and heat fluxes, and the formation of Labrador seawater. The induced change of AMOC may be delayed by years to decades, leading to an ocean-atmosphere coupled oscillation. The time-scale of AMO is determined by the time lag between NAO forcing and THC response (see schematic in Sutton et al., 2018, their Fig. 3).

The zonal-meridional oscillator theory considers internal ocean dynamics (Dijkstra et al., 2006, their Fig. 7). A warm anomaly in the north-central part of the Atlantic basin causes a positive meridional perturbation temperature gradient, which induces via the thermal wind balance a negative zonal surface flow. The anomalous anticyclonic circulation around the warm anomaly causes southward (northward) advection of cold (warm) water to the east (west) of the anomaly, resulting in westward phase propagation of the warm anomaly. Due to this westward propagation, the zonal perturbation temperature gradient becomes negative, inducing a negative surface meridional flow. The resulting upwelling (downwelling) perturbations along the northern (southern) boundary cause a negative meridional perturbation temperature gradient, inducing a positive zonal surface flow, and the second half of the oscillation starts. The time-scale of AMO is determined by the phase difference between the zonal and meridional surface flow perturbations and the westward propagation of the temperature anomalies.

The delayed advective oscillator theory also considers internal ocean dynamics (Lee & Wang, 2010, their Fig. 3). The warm AMO phase with a warmer SST in the subtropics is in favour of generating a decreased meridional density gradient, which in turn leads to a weakened AMOC. The time-scale of AMO is determined by the time delay in the advective flux response to a change in meridional density gradient. This time delay, which is related to the basin-crossing time of long baroclinic Rossby waves at the high-latitude North Atlantic, allows alternating actions of positive and negative advective feedbacks and thus gives rise to a self-sustained oscillation.

The ice coupled oscillator theory considers the interactions among THC, atmospheric winds and Arctic sea ice transport (Dima & Lohmann, 2007, their Fig. 8). An enhanced THC generates uniform positive SST anomalies in the North Atlantic. The atmospheric response is represented by an SLP low that extends over the SSTs but also farther eastward over Eurasia. The multidecadal signal is also transferred in the Pacific basin via the tropics. Further, it affects the Arctic Pacific through atmospheric teleconnections, where it manifests as a weakened Aleutian low and associated positive SST anomalies extending eastward from the East Asian coast. The local positive feedback that involves oceanic adjustment amplifies these structures, which are reaching a maximum amplitude after 10–15 yr. At this time, the SLP structure includes a maximum gradient over Fram Strait, increasing significantly the Arctic sea ice and freshwater export. Consequently, the meridional overturning in the...
ocean is reduced due to resulting freshwater fluxes in the North Atlantic (10-20 yr) and the cycle is turned into its opposite phase. The time-scale of AMO is determined by several time lags in the loop, especially the spinup/down time of the THC.

The cloud-radiation feedback theory does not involve the THC, but considers only interactions between the atmosphere and ocean mixed layer, which is represented by a slab ocean model (Clement et al., 2015, Fig. 4c,d). However, many studies using slab ocean models do not support that the AMO could be generated without involving ocean dynamics (Garuba et al., 2018; Li et al., 2020; Oelsmann et al., 2020; O’Reilly et al., 2016; Sun et al., 2018; Wills et al., 2019; Zhang, 2017; Zhang et al., 2016).

The aerosol forcing theory emphasizes the radiative forcing of human-released and volcano aerosols (Booth et al., 2012, Fig. 4e, f). Aerosol emissions and periods of volcanic activity can explain a large fraction of the simulated multidecadal variance in detrended 1860–2005 North Atlantic sea surface temperatures, which is supported by other comparisons of forced and unforced model experiments (Bellomo et al., 2018; Frankignoul et al., 2017; Murphy et al., 2017; Terray, 2012; Ting et al., 2014; Watanabe & Tatebe, 2019). Birkel et al. (2018) suggested that cool SST patterns develop in association with an increased prevalence of NAO+ patterns caused by stratospheric aerosol loading and a steepened poleward temperature gradient. NAO+ patterns promote wind-driven advection, evaporative cooling, and increased albedo from enhanced Saharan dust transport and anthropogenic aerosols. SSTs across the subpolar gyre are regulated by strength of low pressure near Iceland and the associated wind-driven advection of cold surface water from the Labrador Sea.

The first generation of paleoclimate reanalysis for the Common Era makes it possible to examine with statistical confidence the structure and physical mechanisms of the AMO. Hakim et al. (2016) developed the Last Millennium Reanalysis (LMR) using an “offline” approach to data assimilation, where static ensemble samples are drawn from existing CMIP climate-model simulations to serve as the prior estimate of climate variables. They used linear, univariate forward models (proxy system models) that map climate variables to proxy measurements by fitting proxy data to 2 m air temperature from gridded instrumental temperature data; the linear PSMs are then used to predict proxy values from the prior estimate. Results for the LMR are compared against six gridded instrumental temperature data sets and 25% of the proxy records are withheld from assimilation for independent verification. Results show broad agreement with previous reconstructions of Northern Hemisphere mean 2 m air temperature, with millennial-scale cooling, a multicentennial warm period around 1000 C.E., and a cold period coincident with the Little Ice Age (circa 1450–1800 C.E.). Verification against gridded instrumental data sets during 1880–2000 C.E. reveals greatest skill in the tropics. Tardif et al. (2019) developed LMR version 2 based on an updated proxy database and improved seasonal regression-based forward models. Validation against various instrumental-era gridded analyses shows that the new reconstructions of surface air temperature and 500 hPa geopotential height are significantly improved. Steiger et al. (2018) developed the Paleo Hydrodynamics Data Assimilation product (PHYDA) using a data assimilation approach tailored to reconstruct hydroclimate that optimally combines 2,978 paleoclimate proxy-data time series with the physical constraints of an atmosphere-ocean climate model. The global reconstructions are annually or seasonally resolved and include two spatiotemporal drought indices and near-surface air temperature. Neukom et al. (2019) further derived global surface temperature fields for the past 2000 years using six different reconstruction methods. Similar to the reanalysis of modern instrument data, the paleoclimate reanalysis is affected by data availability. There are more paleoclimate records in the northern hemisphere and over the continents, and the paleoclimate reanalysis output over the southern oceans are more likely coming from the models.

For the external forcings, IPCC AR6 has released solar forcing (Matthes et al., 2017) and volcano forcing (Luo, 2018) datasets for the past 2000 years. Tidal gravitational force can be calculated from the planetary and lunar ephemerides. In astronomy an ephemeris is a tabulation of the positions of a planet or satellite for a series of equally spaced dates, which is designed for astronomical observations and navigation of ships, airplanes and spacecrafts (Seidelmann, 2019). The modern ephemerides are based on dynamical models of the solar system constrained by observational datasets. The first generation of numerical ephemerides were developed in the 1950s–1960s (Ash, 1965; Devine & Dunham, 1966; Eckert et al., 1951). Then numerical ephemerides were developed and continuously improved by several groups around the world, especially for supporting the space missions, such as the Development Ephemerides (DE) by NASA Jet Propulsion Laboratory (Folkner et al., 1994, 2014; O’Handley et al., 1969; Standish, 1982, 1990; Standish et al., 1976), the Ephemerides of Planets and the Moon (EPM) by the Institute of Applied Astronomy of Russia (Akim & Stephanian, 1977; Krasinsky et al., 1981, p. 1982; Akim et al., 1986; Pitjeva, 2005, 2013), and the Intégration Numérique Planétaire de l’Observatoire de Paris (INPOP) by the IMCCE-Observatoire de Paris (Fienga et al., 2008, 2019). These ephemerides combine the best theories and are fitted to optical, radar and increasingly dense sets of space mission data. Their dynamical model includes point-mass interactions between the nine planets, the Sun and asteroids, relativistic PPN effects (Moyer, 1971, 2000), Fig. effects, Earth tides and lunar librations (Newhall et al., 1983). Major improvements in observational accuracy led by modern technology, such as Lunar Laser Ranging, range and VLBI spacecraft tracking, have led to comparable improvements in the accuracy of the planetary and lunar ephemerides. IMCCE also developed analytical theories called the Variations Séculaires des Orbits Planétaires.
Fig. 4  Regression of Atlantic basin zonal mean (a) temperature, and (b) salinity to THC index at lag +10 years (from Delworth et al., 1993). Regression of SST (shaded), SLP (contours), and surface winds (vectors) to the AMO index for (c) fully coupled models, and (d) slab-ocean models (from Clement et al., 2015). Differences between warm and cold AMO phases for (e) aerosol burden, and (f) surface net shortwave flux (from Booth et al., 2012).
(VSOP) and the Theory of the Outer Planets (TOP), which were fitted to the above numerical ephemerides (Bretagnon & Francou, 1988; Simon et al., 2013). The ephemerides generally cover the past several thousand years.

A long-lasting question in AMO research is if the AMO is an oscillation or a variability of a broad spectral band. The modern SST analysis covers only 2.5 AMO cycles and thus show a broad peak centered at 60 years (not shown). Fig. 5 illustrates the maximum entropy spectra for the past 2000 years of (a) North Atlantic Ocean SST from LMR reanalysis (AMO index as defined by Enfield et al., 2001), (b) Nino3.4 SST from LMR reanalysis, (c) global mean surface temperature from LMR reanalysis, (d) AMOC index from Mjell et al. (2016), (e) solar forcing from IPCC AR6, (f) volcano forcing from IPCC AR6, and (g) tidal gravitational forcing calculated from NASA JPL DE431. The AMO index demonstrates two well-defined spectral peaks near 60 years (Fig. 5a). Other paleoclimate reanalysis datasets also show similar results. The Nino3.4 SST anomaly (Fig. 5b) and global mean surface temperature (Fig. 5c) from the LMR reanalysis demonstrate strong spectral peaks between 40–100 years, which are confirmed by other paleoclimate reanalysis datasets (not shown). The independent AMOC index from Mjell et al. (2016) shows strong spectral peaks between 40–100 years (Fig. 5d), suggesting that the AMO is associated with deep ocean thermohaline circulation, and correlation analysis will be shown below to confirm the relationship. Regarding the external forcings, the solar forcing does not have any significant multi-decadal variability (Fig. 5e). The volcano forcing (Fig. 5f) and tidal gravitational forcing (Fig. 5g) both show spectral peaks between 40–100 years, and correlation analysis will be shown below to confirm their relationships with the AMO. Therefore, the AMO is a well-defined oscillation rather than a red-noise-like variability, and is a global phenomenon, whose spatial structure will be illustrated below.

Fig. 6 illustrates the global structure of the AMO throughout its lifecycle, which was not possible to be derived from the short modern instrument data. This Fig. shows the lag-correlation of 40–100-year filtered LMR reanalysis SST anomaly with AMO index for lag (a) -25 years, (b) -20 years, (c) -15 years, (d) -10 years, (e) -5 years, and (f) 0 year. Stars denote the grids with correlation coefficients above the 95% confidence level. At lag -25 years (Fig. 6a), the global SST pattern is opposite to the AMO SST pattern from modern instrument data as discovered by Enfield et al. (2001) (Fig. 1b). There are significant cold SST anomalies in North Atlantic, tropical Atlantic, tropical Indian Ocean, northwest Pacific, northeast Pacific, and southeast Pacific. There are regions free of significant SST anomalies, such as the equatorial central and eastern Pacific, north Pacific and south Pacific. At lag -20 years (Fig. 6b), the cold SST anomalies start to weaken, while the regions free of significant SST anomalies start to show some warm SST anomalies. At lag -15 years (Fig. 6c), the cold SST anomalies almost disappear, while the regions free of significant SST anomalies in Fig. 6a are occupied by significant warm SST anomalies. The pattern is similar to the SST pattern of a weak El Nino. At -10 years (Fig. 6d), the warm SST anomalies grow wider and also occupy North Atlantic, tropical Atlantic and tropical Indian Ocean, which is similar to the SST pattern one season after the peak of El Nino. The SST pattern grows stronger at -5 years, although SST anomaly in equatorial central Pacific starts to decay (Fig. 6e). At 0 year, the North Atlantic SST anomaly reaches its peak, while the equatorial central/eastern Pacific SST anomalies disappear (Fig. 6f).

To understand the physical mechanisms of the AMO, we first examine the relationship between the AMO and the AMOC. Fig. 7 illustrates the lag-correlation of 40–100-year filtered LMR reanalysis SST anomaly with AMOC index from Mjell et al. (2016) for lag (a) -15 years, (b) -10 years, (c) -5 years, (d) 0 year, (e) +5 years, and (f) +10 years. The two datasets are independent of each other. The Mjell et al. (2016) AMOC index comes from measurements of deep ocean currents, which was not assimilated into the LMR reanalysis. This tends to decrease the correlations between the AMOC index and the LMR reanalysis SST. Nevertheless, the global SST anomalies show significant correlations with the AMOC index, while the evolution of the global SST anomalies associated with the AMOC index is similar to those associated with the AMO index (Fig. 6), suggesting that the AMO is connected to the AMOC. It is important to note the 10-year phase difference between Fig. 7 and Fig. 6, which suggests that the strongest AMOC flow leads the warmest North Atlantic SST by about 10 years. At the time of the strongest AMOC flow (Fig. 7d), the warm SST anomalies in the Atlantic Ocean are symmetric about the equator with no significant anomaly right on the equator, suggesting ocean upwelling near the equator and poleward advection at the surface in both hemispheres. Then the warm SST anomalies in North Atlantic expand further northward and occupy most of the North Atlantic Ocean (Fig. 7e,f). Therefore, the AMO definitely involves the oscillations of the AMOC as proposed by the ocean-related AMO theories such as the stochastic forcing theory, the NAO coupled oscillator theory, the delayed advective oscillator theory and the ice coupled oscillator theory.

For the aerosol forcing theory, Fig. 8 demonstrates the lag-correlation of 40–100-year filtered LMR reanalysis SST anomaly with the volcano forcing from IPCC AR6 for lag (a) 0 year, (b) +5 years, and (c) +10 years. At lag 0 year, volcano eruptions are associated with cold SST anomaly in North Atlantic, Arctic, South Atlantic, North Pacific and equatorial western Pacific oceans (Fig. 8a), which is likely caused by the reflection of sunlight by volcano dust. The cold SST anomalies grow at lag +5 years (Fig. 8b) and persist to lag +10 years (Fig. 8c). These results suggest that
the volcano forcing enhances the AMO negative phase as proposed by the aerosol forcing theory.

For the cloud-radiation feedback theory, Fig. 8 also illustrates the lag-correlation of 40–100-year filtered LMR reanalysis surface downward shortwave flux anomaly with AMO index for lag (d) -10 years, (e) -5 years, and (f) 0 year. The surface downward shortwave flux anomalies are mainly controlled by the clouds. Note that the direction of
The flux is downward and positive (negative) anomaly will warm up (cool down) the SST anomaly. Comparing the surface downward shortwave flux anomalies (Fig. 8d–f) with the corresponding SST anomalies (Fig. 6d–f) indicates that the surface downward shortwave flux provides negative feedback in the North Atlantic at lag -10 years, which gradually changes to positive feedback at lag -5 years and lag 0 year. The sign of the feedback varies in other oceans. For example, the feedback is largely negative in equatorial central Pacific. Therefore, cloud-radiation feedback enhances the AMO in North Atlantic at its peak time as proposed by the cloud-radiation feedback theory. However, cloud-radiation feedback suppresses the SST anomalies at other time or in other ocean basins.

For the tidal gravitational forcing, Fig. 9 demonstrates the lag-correlation of 40–100-year filtered LMR reanalysis SST anomaly with tidal gravitational force calculated from NASA JPL DE431 ephemeris for lag (a) -20 years, (b) -10 years, (c) -5 years, (d) -1 years, (e) 0 year, and (f) +5 years. The tidal gravitational force is highly correlated with the SST anomalies.
Since the tidal gravitational force is totally independent of the LMR reanalysis, it is amazing to see such significant correlations. Comparison between Fig. 9 and Fig. 6 shows that the evolution of the global SST anomalies associated with the tidal gravitational force is similar to those associated with the AMO index, suggesting that tidal gravitational force plays a key role in driving the AMO. Note the 10-year phase difference between Fig. 9 and Fig. 6, which indicates that the strongest tidal force leads the warmest North Atlantic SST by about 10 years, and is in phase with the strongest AMOC flow (Fig. 7), suggesting that tidal gravitational force drives the AMOC flow consistent with the AMOC theories (Kuhlbrodt et al., 2007). At the time of the strongest tidal gravitational force (Fig. 9d), the warm SST anomalies in the Atlantic Ocean are symmetric about the equator with weak or no significant anomaly right on the equator, suggesting ocean upwelling near the equator and poleward advection at the surface in both hemispheres. Then the warm SST anomalies in North Atlantic expend further northward and occupy most of the North Atlantic Ocean (Fig. 9e,f). There is a persistent horseshoe-shape cold
SST anomaly in the Pacific Ocean in Fig. 9e,f which has been replaced by warm SST anomalies in Fig. 6 and Fig. 7, suggesting other physical mechanisms contribute to the warm SST anomalies in that region in Fig. 6 and Fig. 7.

Fig. 10 summarizes schematically the global SST structure and physical mechanisms of the AMO. The AMO is a global-scale coupled ocean-atmosphere oscillation of the climate system with significant SST anomalies in all ocean basins. It is associated with significant oscillation of the AMOC flow, which is likely driven by tidal gravitational forcing and enhanced by volcano forcing and cloud-radiation feedback. Tidal mixing has been found to be a leading physical mechanism for exciting the AMOC (Kuhlbrodt et al., 2007 and other references cited before), as well as upper ocean circulation (Hasumi et al., 2008; Tanaka et al., 2012; Osafune & Yasuda, 2013; Osafune et al., 2014, p. 2020; Tatebe et al., 2018). Therefore oscillations in tidal forcing are likely causing oscillations of the AMOC, global thermohaline circulation and upper ocean circulation. Tidal gravitational force is also known to drive global tectonics (Bendick & Bilham,
which may be linked to the multi-decadal variability of volcano activity.

3 AMO teleconnections and impacts

The major difficulty for studying AMO teleconnections and impacts is the short length of modern instrument data, which cover only two and a half AMO cycles. Although coherent oscillations of physical variables were often found, it is difficult to get statistically significant correlations with AMO index for key fields such as the sea level pressure and land surface air temperature (Fig. 11a,b, Alexander et al., 2014), or monsoon precipitation (Fig. 11(c,d); Zhang & Delworth, 2006; Wang et al., 2009). Other examples include the Hadley circulation (Guo et al., 2016; Stuckman, 2016), northern hemisphere climate modes (Li et al., 2018; Wyatt et al., 2012), and tropical cyclone activity (Bell & Chelliah, 2006; Caron et al., 2015; Goldenberg et al., 2001; Klotzbach, 2011; Klotzbach & Grey, 2008; Knight et al., 2006; Zhang et al., 2018; Zhang & Delworth, 2006; see review by Lin et al., 2022) Grey.

Fig. 9 Lag-correlation of 40–100-year filtered LMR reanalysis SST anomaly with tidal gravitational forcing from NASA JPL for lag (a) -15 years, (b) -10 years, (c) -5 years, (d) 0 year, (e) +5 years, and (f) +10 years. Stars denote the grids with correlation coefficients above the 95% confidence level.
Many studies have used global climate models to understand the AMO teleconnections and impacts (Feng & Hu, 2008; Goswami et al., 2006; Guan & Nigam, 2009; Kucharski et al., 2015; Li et al., 2008; Rashid et al., 2010; Ruprich-Robert et al., 2017; Timmermann et al., 2007; Yu et al., 2015; Zhang & Delworth, 2005). Using a coupled ocean-atmosphere model, Zhang and Delworth (2005) found that a large reduction in the Atlantic thermohaline circulation (THC) can induce global-scale changes in the tropics (Fig. 12). The weakening of the Atlantic THC causes a southward shift of the intertropical convergence zone over the Atlantic and Pacific, an El Nino-like pattern in the southeastern tropical Pacific, and weakened Indian and Asian summer monsoons through air–sea interactions.

The impacts of AMOC on global surface temperature are more complicated. Recently there was a debate between Chen and Tung (2018, 2021) and Caesar et al. (2020; 2021) on whether the rate of global warming is enhanced by a weak AMOC.

The first generation of paleoclimate reanalysis for the past 2000 years makes it possible to examine with statistical confidence the global teleconnections and impacts of the AMO. For global teleconnections, we analyzed the SLP and 500 mb geopotential height (Z500). Fig. 13 illustrates the lag-correlation of 40–100-year filtered LMR reanalysis SLP anomaly with AMO index for lag (a) -30 years, (b) -15 years, and (c) 0 year. At lag -30 years, there are surface high pressure anomalies over the Arctic Ocean, central Asia, tropical Africa and the southern oceans, but low pressure anomalies over Europe, south Atlantic, equatorial central Pacific and Ross Sea (Fig. 13a). A quarter cycle later at lag -15 years, there are high pressure anomalies over Antarctica and North America, but low pressure anomalies over most of the Pacific Ocean and the southern oceans (Fig. 13b). The SLP pattern at lag 0 year (Fig. 13c) mirror that at lag -30 year (Fig. 13a).

For Z500, at lag -30 years, most of the Northern Hemisphere are covered by negative Z500 anomalies (Fig. 13d), suggesting that there are cold temperature anomalies in the lower troposphere. At lag -15 years, there are positive Z500 anomalies over Antarctica and Aleutian Islands, but negative Z500 anomalies over the southern oceans and Hawaii (Fig. 13e). At lag 0 year, most of the Northern Hemisphere are covered by positive Z500 anomalies (Fig. 13f), suggesting that there are warm temperature anomalies in the lower troposphere.

For the global impacts of AMO, we examined the surface air temperature (Tair) and Palmer Drought Severity Index (PDSI) over the continents. The AMO strongly affects surface air temperature over the continents (Fig. 14). At lag -30 years (Fig. 14a), most of the global continents are covered by cold Tair anomalies. The cold anomalies weaken at lag -20 years (Fig. 14b), change to warm anomalies at lag -10 years (Fig. 14c), and reach the peak at lag 0 year (Fig. 14d). The warm anomalies wither at lag +10 years (Fig. 14e), and switch to cold anomalies at lag +20 years (Fig. 14f), which continue the cycle to lag -30 years (Fig. 14a).
Fig. 11 Correlation of AMO index with (a) sea level pressure and (b) surface air temperature for NOAA 20C reanalysis (1871-2008). Shading denotes regions with correlation above 95% confidence level (from Alexander et al., 2014). (c) Regression of JAAS precipitation in Africa and India to AMO index for 1900–2002 (from Zhang & Delworth, 2006). (d) Regression of JJA precipitation in Asia to AMO index for 1900-2002. Green contour denotes regions above 95% confidence level (from Wang et al., 2009).
The AMO also strongly affects the multi-decadal mega-droughts around the world (Fig. 15). With the 40–100-year oscillations, the dry/wet periods last for 20–50 years. Note that negative values of PDSI mean droughts. For example, at lag -30 years (Fig. 15a), there are droughts over tropical Africa, but wet anomalies over Europe, south Africa,
Siberia, Canada, Brazil and Peru. At lag -10 years (Fig. 15c), there are droughts over Canada, south Africa, Australia and Brazil, which amplify and expand at lag 0 year (Fig. 15d). The PDSI patterns are more complex than the Tair patterns (Fig. 14), since the PDSI is affected not only by Tair, but also by convection, land surface processes and other factors.

Fig. 16 summarizes schematically the global impacts of the AMO. Red (blue) shading denotes the region with statistically significant surface air temperature (PDSI) anomalies. The
phase lag to the AMO index is represented by the arrows. The phase clock is also shown. The AMO strongly affect $T_{\text{air}}$ and PDSI around the world. The phase lag varies among different continents. Over the tropical continents, $T_{\text{air}}$ and PDSI tend to be in phase with each other. Over the extratropical continents, $T_{\text{air}}$ and PDSI tend to be out of phase with each other.
4 AMO and the global warming hiatus

The global surface temperature warming paused during 2000–2012, which happened over both land and oceans and was termed the “global warming hiatus” (Knight et al., 2009; Easterling & Wehner, 2009; Liebmann et al., 2010; Wang et al., 2010; Foster & Rahmstorf, 2011; Santer et al., 2011, 2012; IPCC, 2013; Trenberth & Fasullo, 2013; Fyfe et al., 2013). There were debates on the detailed definition of
global warming hiatus was associated with a mysterious “missing energy” in Earth’s energy budget. Earth’s radiation budget can be accurately measured by satellites (Kyle et al., 1985; Ramanathan et al., 1989; Ardanuy et al., 1991; Kiehl et al., 1994; Wielicki et al., 1995; Loeb et al., 2009; Kiehl & Trenberth, 1997; Trenberth et al., 2009). The Earth’s climate system had a net energy gain of 0.5-1 Wm⁻² in the first decade of the twenty-first century (Allan et al., 2014; Balmaseda et al., 2013; Hansen et al., 2011; Loeb et al., 2012; Trenberth et al., 2014), which was caused mainly by the heating due to continuously increasing greenhouse gases after partially offset by cooling due to a low solar minimum (Hansen et al., 2011), decrease in stratosphere water vapour (Solomon et al., 2010), multiple small volcano eruptions (Vernier et al., 2011), and increasing anthropogenic aerosols (Kaufmann et al., 2011). This net energy gain was confirmed by both the satellite measurements of radiation fluxes at the top of the atmosphere and all the different estimates of global ocean heat content (OHC; Balmaseda et al., 2013; Ishii & Kimoto, 2009; Katsman & van Oldenborgh, 2011; Levitus et al., 2009, 2012; Liu & Xie, 2018; Palmer et al., 2011; Watanabe et al., 2013; X. Hu et al., 2018a).

However, the global surface temperature warming paused during this period, which happened over both the land and the ocean. Then where has the energy gain gone inside the Earth’s climate system? Among the different components of the Earth’s climate system, the atmosphere, land and cryosphere all have small heat capacity, while the ocean stores about 90% of the total heat content (Balmaseda et al., 2013). Observational analysis of global OHC revealed that the missing energy has been sequestered into the global deep ocean below 700 m deep, while the upper ocean (0–700 m) got little heat and had a hiatus similar to the ocean surface (Balmaseda et al., 2013; Levitus et al., 2009, 2012; Trenberth et al., 2014).

Observational and modelling studies have attributed the global warming hiatus to the Inter-decadal Pacific Oscillation (Clement & DiNezio, 2014; Douville et al., 2015; England et al., 2014; Kosaka & Xie, 2013, 2016; Meehl et al., 2013, 2014; Watanabe et al., 2014; Deser et al., 2017; Wang et al., 2017a), Pacific Decadal Oscillation (Dai et al., 2015; Trenberth, 2015), ENSO (Hu & Fedorov, 2017), Indian Ocean (Lee et al., 2015; Nieves et al., 2015), and the AMO (Chen & Tung, 2014; Chylek et al., 2014; DelSole et al., 2011; Kravtsov et al., 2015; Mann et al., 2014; McGregor et al., 2014; Steinman et al., 2015a, 2015b; Thompson et al., 2009; Tung & Zhou, 2013; Wu et al., 2007, 2011; Yao et al., 2016; Zhang et al., 2007; Zhou & Tung, 2013; Haustein et al., 2019; Hu et al., 2018b). Meehl et al. (2013) used the Community Climate System Model version 4 (CCSM4) to examine the physical mechanisms of global warming hiatus in the model. They found that accelerated warming decades are characterized by rapid warming of globally averaged surface air temperature, greater increases of heat content in the upper ocean layers, and less heat content increase in the deep ocean, opposite to the hiatus decades. In addition to contributions from processes potentially linked to Antarctic Bottom Water (AABW) formation and the AMOC, the negative phase of the IPO, adding to the response to external forcing, is usually associated with hiatus decades. Wu et al. (2007) showed that the rapidity of the warming in the late twentieth century was a result of concurrence of a secular warming trend and the warming phase of a multidecadal (~65-year period) oscillatory variation and they estimated the contribution of the former to be about 0.08 °C per decade. Chen and Tung (2014) used in situ and reanalyzed data to trace the pathways of ocean heat uptake. In addition to the shallow La Niña–like patterns in the Pacific that were the previous focus, we found that the slowdown is mainly caused by heat transported to deeper layers in the Atlantic and the Southern oceans, initiated by a recurrent salinity anomaly in the subpolar North Atlantic. Cooling periods associated with the latter deeper heat-sequestration mechanism historically lasted 20–35 years.

The first generation of paleoclimate reanalysis provides the global mean surface temperature for the past 2000 years. PAGES 2k Consortium (2019) presented the global mean surface temperature for the Common Era based on seven reconstruction methods and found strong consistent multi-decadal oscillations (Fig. 17b). The maximum entropy spectrum shows strong peaks around 60 years (Fig. 5c), which is consistent with the maximum entropy spectrum of the global mean surface temperature from modern instrument records (Fig. 17c). Fig. 18 illustrates...
Fig. 17  (A) Observed (HadCRUT4.6, NOAAGlobalTemp) and CMIP6 model global mean surface temperature change for 1850-2020. Shading shows ±1 standard deviation about the ensemble mean (from Modak & Mauritsen, 2021). (B) Observed temperature anomalies for the past 2000 years from seven paleoclimate reconstruction methods together with CMIP5 models subjected to 30–200 yr bandpass filters (from PAGES 2k Consortium, 2019). (C) Maximum entropy spectra of observed global mean surface temperature from three datasets for 1880-2016.
Fig. 18  Lag-correlation of 40–100-year filtered LMR reanalysis SST anomaly with global mean surface temperature anomaly for lag (a) -25 years, (b) -20 years, (c) -15 years, (d) -10 years, (e) -5 years, and (f) 0 year. Stars denote the grids with correlation coefficients above the 95% confidence level.
the lag-correlation of 40–100-year filtered LMR reanalysis SST anomaly with global mean surface temperature anomaly for lag (a) -25 years, (b) -20 years, (c) -15 years, (d) -10 years, (e) -5 years, and (f) 0 year. Stars denote the grids with correlation coefficients above the 95% confidence level. The SST pattern and evolution are similar to those of the AMO (Fig. 6). Therefore, the observed multi-decadal oscillation of global mean surface temperature is caused by the AMO. The ocean subsurface variables are not available from current paleoclimate reanalysis. The SODA ocean reanalysis using modern instrument data shows that the AMO is associated with a temperature dipole between the upper ocean (0–700 m) and the deep ocean (700m – bottom), while the IPO is associated with a temperature dipole within the upper ocean between 0–150 m and 150–700 m (not shown). These results suggest that the heat sequestration into deep ocean below 700 m was caused by the AMO, not the IPO.

5 Summary and future directions

This paper reviews the history of AMO research on its dynamics, global teleconnections and impacts, and connection to global warming hiatus. The AMO is a global-scale coupled ocean-atmosphere oscillation of the climate system with significant SST anomalies in all ocean basins. It is associated with significant oscillation of the AMOC flow, which is likely driven by tidal gravitational forcing through tidal mixing and enhanced by volcano forcing and cloud-radiation feedback. The AMO strongly affects Tair and PDSI over all the continents. The phase lag varies among different continents. Over the tropical continents, Tair and PDSI tend to be in phase with each other. Over the extratropical continents, Tair and PDSI tend to be out of phase with each other. The AMO contributes significantly to global mean surface temperature and global warming hiatus.

The long-lasting bottleneck of AMO research was the short length of modern instrument records. The first generation of paleoclimate reanalysis has helped to break this bottleneck. The first generation of paleoclimate reanalysis provides surface variables and Z500. We look forward to seeing paleoclimate reanalysis of the global ocean subsurface temperature, salinity and currents, which will make it possible to analyze global ocean circulation structure associated with the AMO. More upper air variables would also make it possible to analyze the AMO impacts, such as those on the advance/retreat of mountain glaciers.

The AMO switched to its cooling phase around 2004 (e.g. Chen & Tung, 2018). If AMO keeps its 50-70-year timescale, it will likely extend the recent global warming hiatus to the 2030s, which is similar to what happened before in 1884–1911 and 1943–1974 (Fig. 17a), and will largely mask the centennial global warming. After AMO switches to its warming phase in the 2030s, it may enhance the global warming thereafter. Therefore, it is important to forecast the AMO, so that the human society could watch for the possible forthcoming reduced/enhanced global warming periods.

Include tidal gravitational forcing into climate models may help to improve their AMO forecasts and thus decadal climate predictions. Recently, the ocean modelling community show strong interest in lunar tidal forcing because of the discovery that tidal mixing plays a key role in global ocean circulation (Loder & Garrett, 1978; Melet et al., 2016; Schmittner et al., 2015). Parameterizations of diurnal and semidiurnal tidal mixing have been implemented into several OGCMs such as the GFDL MOM (Schiller & Fiedler, 2007), HYCOM (Arbic et al., 2010), and MIROC (Tanaka et al., 2012). However, for simulating the multi-decadal tidal components related to AMO, explicit modelling of time-varying gravitational field is needed. The MPI OM group has developed a tidal forcing option to include explicit time-varying gravitational forcing from the Sun and Moon including the seasonal, annual, interannual and inter-decadal tidal cycles (Muller et al., 2012). For each time step of simulation, the actual positions of the Sun and Moon are calculated using the semi-analytic planetary theory Variations Seculaires des Orbites Planetaires (VSOP87; Bretagnon & Francou, 1988), and the associated gravitational forcing is determined. This tidal forcing option has not been used in the MPI model’s climate predictions (Baehr et al., 2015) or IPCC runs (Muller et al., 2018). Nevertheless, the MPI model has demonstrated that it is possible to add to GCMs explicit time-varying gravitational forcing from the Sun and Moon. If the model experiments confirm that lunar tidal forcing drives the AMO through tidal mixing for AMOC, this key new physics will provide valuable long-range predictability, and help to improve the decadal to multi-decadal predictions of global climate change (Cassou et al., 2018; Kirtman et al., 2013; Marotzke et al., 2016; Meehl et al., 2009, 2010).

The Earth’s climate system follows the same physical laws and often involves similar physical processes at different time-scales. Success in understanding and predicting AMO may help understanding and predicting other important climate modes such as the centennial-scale variability, millennial-scale variability and inter-glacial cycles, all of which have been suggested to be connected with variations of AMOC.

Acknowledgments

Part of this work was done by TQ when working in the Department of Geography, The Ohio State University.

Disclosure statement

No potential conflict of interest was reported by the author(s).
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