Title
Atmospheric and Oceanic Circulation Patterns

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Abstract
Fueled by the Sun, intrinsic dynamics of Earth systems drive global climate. Multiple interactions among climate-system components – ocean, atmosphere, land, and ice – are influenced by geography. They are further complicated by external forcings, both natural and anthropogenic. Dissimilar response times of system-components to perturbations, combined with non-linear reactions among the components, have potential to generate oscillatory signals on a variety of timescales. Positive and negative feedbacks further complicate these manifold interactions. Complex interplay among these various processes results in redistribution of planetary heat, basically from where it is plentiful to where it is not. Through process-related changes in wind and precipitation regimes; atmospheric chemistry; the biosphere; inventories of clouds, sea-ice, snow-cover, and the like, Earth's radiative-energy balance – the difference between incoming and outgoing energy – is modulated. These climate-driving dynamics can be distilled into a global collection of individual oceanic and atmospheric circulation patterns. In the short-term, these patterns manipulate local and remote weather. Cumulatively and collectively, in the long-term, they influence climate, their combined effect on global surface temperature complex.

Introduction:

Climate: the Big Picture

Weather is a cooling process. The Sun heats Earth's surface; weather removes the surplus. Weather, averaged over time, is climate. Climate is what happens between delivery of energy and exit of its excess, the net-result of which modulates Earth's average surface temperature.

The details are less simple. Interactions among a myriad of internal and external boundary conditions result in the absorption and partitioning of heat among various reservoirs – ocean, ice, land, and atmosphere – and at various levels within each. On different time scales, from seasonal to millennial, stored heat is moved in an assortment of ways via processes, many of which are coupled, to new latitudes, longitudes, depths, altitudes, and even to different reservoirs, with the end-result being a global relocation of heat, from where there is an abundance to where there is less. Along this journey of heat re-distribution, some heat escapes, emitted to space. How incoming absorbed (solar) energy matches up to outgoing emitted infrared (heat) energy determines Earth's radiative energy balance. Climate processes collectively effect, over time, the uptake, re-distribution, and exit of planetary heat. Radiative-energy imbalances are thereby minimized, their influence on surface temperature complex.

Climate's Task Masters: Atmospheric and Oceanic Circulation Patterns

Processes whose collective outcome modulates Earth's radiative-energy imbalances can be traced to a global collection of individual oceanic and atmospheric circulation patterns. Locally and regionally, each circulation pattern impacts short-term weather. Interactions between and among them manipulate long-term climate. Each pattern possesses distinct traits.

All patterns fluctuate, most with preferred timescales of variability. A pattern's timescale of variability need not be governed by a similarly cadenced external source. The source of variability can be intrinsic to the system. Different response times of components to a forcing or disturbance, and non-linear interactions (where cause-and-effect relationships are not strictly proportional) among those components, leave the subsystem in a state of disequilibrium. Hence, the components never achieve collective stability. Therefore, they are varying constantly. The complexity of the subsystem, with extreme sensitivity to details regarding perturbations and boundary conditions, ensures that the resulting oscillatory behavior is not rigidly periodic. Feedback responses add another layer of complexity. Due to dissimilar response times of subsystem components to disturbances, it is not unusual for the initial system-response to involve an immediate positive feedback (enhancing the initial perturbation), followed by a delayed negative feedback (damping or reversing the initial response), thereby modifying timescale of variability through this means, as well. And because the fluctuations are not temporally regular, the oscillatory character is best described as quasi-periodic. Most circulation patterns vary at more than one “quasi-regular” frequency, often including a low-frequency component. Preferred frequencies can be internally generated or externally forced, or internally generated with excitation from or entrainment by an external source.

Characteristic processes and traits tend to repeat for a given pattern; yet, again due to the non-linear nature of the subsystems, no two snapshots in time will be identical. In addition, teleconnections – climate features or responses highly correlated to, and distantly forced by, an associated regional circulation pattern - may co-exist with a given pattern for years and then wane or disappear, only to return to its former relationship at a later date – a further reminder of climate’s complexity.

Debate exists over whether or not climate, with its numerous interacting circulation patterns, is itself, chaotic. In the mathematical sense, deterministic chaos refers to behavior that is acutely sensitive to initial conditions. Slight differences in initial conditions result in vastly divergent outcomes in chaotic systems. Were one to know with certainty each element of initial conditions, outcome could be predicted. While this knowledge is unattainable, a chaotic system is largely, but not entirely, unpredictable. Nor is the system without order. While a complex, non-linear system can be chaotic, and often is; it can also merely reflect behavior that appears chaotic – hence the debate. Weather tends to be chaotic. Climate, on the other hand, with its longer timescale, global energy-balance modulation over time, and negative feedbacks buffering the system from extreme disequilibrium, makes the situation less clear. Chaotic or not, it is clear that the climate system is complex and non-linear. It involves components that behave chaotically, but it is not without order.

On multidecadal timescales, individual expression of regional patterns may yield to collective interaction (network behavior), as suggested by recent research [1]. According to this hypothesis, synchronization and local coupling give rise to a low-frequency signal that propagates across the Northern Hemisphere through a sequence of atmospheric and lagged oceanic circulations (Fig. 1). In turn, this collective, slowly varying signature furnishes the modulating low-frequency background upon which higher frequency behavior of regional circulations is superimposed.

Over 30 large-scale modes, or leading patterns, of climate variability have been identified across the globe. Most modes are spatially local or regional; yet with many of these, their impacts are hemispheric to global. Some patterns are dominated by atmospheric processes, some by oceanic ones. Some involve sea ice. Coupling between systems (ocean, ice, and atmosphere) and between levels within a system is common, particularly for low-frequency oscillations. Dizzying detail of the numerous circulation modes renders their complete description unfeasible in limited space. Thus, the goal here is to weave together a brief overview of select individual regional circulation patterns. Discussion touches on their geographically governed dynamics, their individual behaviors, and their interconnectivity. The abridged inventory of patterns described here is divided into two broad categories: circulation patterns of the Tropics and those of the Extratropics. Links to websites featuring graphics related to many of the described patterns can be found after the bibliography at the end of this contribution.

Circulation Patterns of the Tropics:

Geographically Governed Dynamics:

Three tropical-specific traits play large roles in interrupting zonal symmetry imposed along the equator by the annual-mean delivery of solar radiation: 1) At the equator, the Coriolis Effect (the apparent deflection of moving objects due to planetary rotation – a function of latitude) is zero. On either side of the equator, directions of deflection are opposite. 2) Large-scale winds from each hemisphere converge near, but not necessarily aligned with, the geographical equator. Convergence is marked by a region of intense atmospheric convection known as the Intertropical Convergence Zone (ITCZ). And 3) planetary-scale subsurface ocean waves (eastward traveling Kelvin and westward propagating Rossby), when excited by anomalous winds, transmit information regarding overlying atmospheric disturbances zonally across the equatorial basin. Within months (the exact time involved dependent upon basin width), the equatorial subsurface adjusts to overlying atmospheric shifts. In turn, modified subsurface dynamics affect surface conditions. Ocean adjustment time – the ocean's 'dynamic memory' - is shorter in the tropics than elsewhere on the planet. Two aspects, unique to the tropics, make possible this rapid oceanic adjustment to overlying atmospheric anomalies: One: latitudinally confined, eastward-directed Kelvin waves travel swiftly. They exist only along the equator. And two: while westward propagating Rossby waves appear at all latitudes, their velocity is fastest in the near-equatorial zone.

Together, this collection of tropical-specific properties generates a mean state of zonal asymmetry in sea-surface temperatures (SST), subsurface thermocline depths, and atmospheric patterns, and from which zonal and meridional deviations occur with relative rapidity.

Zonal (east-west) Modes of Coupled Ocean-Atmosphere Variability in the Tropics:

El Nino-Southern Oscillation (ENSO):

The mean state of the tropical Pacific is characterized by warm SSTs, a deep thermocline, and strong overlying atmospheric convection in the west; a cool tongue of SSTs and a shallow thermocline in the east. Strong easterly winds co-occur with this zonally non-uniform profile. El Nino Southern Oscillation (ENSO), a coupled oceanic-atmospheric system, fluctuates inter-annually from this mean state. Zonal asymmetry is either magnified (La Nina) or diminished (El Nino).

At the helm of the oceanic-atmospheric coupled system is the sea-level-pressure (SLP) configuration of an equatorially circumnavigating phenomenon, the Southern Oscillation (SO). Zonal exchange of atmospheric mass along the equator manifests as fluctuations in the normalized difference in SLP anomalies between high-and low-pressure-centers of the SO. SLP differences between a low-pressure region over north central Australia (near Darwin, Australia (12º28'S; 130º51'E)) and a high-pressure zone over the central South Pacific (Tahiti (17º40'S; 149º25'W)), are reflected in the SO index (SOI), and is one metric used to define ENSO phase and intensity. A negative SOI describes the ENSO warm phase (El Nino), a positive SOI a cool one (La Nina). Surface waters slosh east-west in concert with the SOI, influencing subsurface and overlying features in their wake. Adjustments of the subsurface ocean and overlying atmosphere to the equatorial zone's SLP redistribution occur regionally. Their impact is global.

Positive and negative departures from the tropical Pacific mean state are initially magnified through positive feedback responses (Bjerknes mechanism [2]). Interplay of subsurface Kelvin and Rossby ocean waves initially amplifies, then ultimately damps, or even reverses sign of, the excursions. Together, the Bjerknes mechanism and the lagged Kelvin-Rossby interaction describe a leading model for the ENSO pattern, a delayed oscillator system [3-5].

While El Nino events rarely conform to a typical profile, a 'standard' or 'canonical' El Nino can be described. Hints of a building 'canonical' El Nino can be seen in decreased SLP over Tahiti and increased SLP over Darwin. With the corresponding drop in SLP difference between the two pressure-centers, southeasterly winds over the south and central tropical Pacific weaken. The atmospheric 'push' on warm waters pooled in the west relaxes. Warm waters spread eastward. Westerly wind anomalies further advance the collapse of zonal asymmetry along the equatorial zone. Initial signs of an El Nino event typically manifest in late fall, particularly around the end of December (thus its name "El Nino" (Christ child) in reference to the time of year). Once started, an incipient El Nino may reverse course and disappear. But if it survives seasonally competing processes imposed on its development during the boreal spring (e.g. seasonally related strengthened southeasterlies), it generally matures. El Nino 'onset' typically takes hold around March, often persisting 18 to 24 months thereafter.

With El Nino's continued evolution, a weakened atmospheric zonal component (Walker Circulation) shifts eastward with the eastward spread of warm SSTs and precipitation. Dry conditions supplant the moisture displaced from the western tropical Pacific. Upper-level subtropical jets on either side of the equator strengthen and shift equatorward, in particular the jet within the winter hemisphere. Jet-flow is extended eastward. Weather systems are re-directed accordingly. Regions may experience warmth where usually it is cool; moisture where usually it is dry.

Teleconnection features associated with a circulation pattern can persist for decades and then wane or disappear. An example lies in the strong correlation between El Nino events and the suppression of North Atlantic hurricanes. Such was the norm for much of the 1950s into the 1980s. In the mid-1980s, and throughout the 1990s, this relationship began to falter. Spatial distribution and location of El Nino-related SST warming play a role. While warm SST anomalies in the eastern basin characterized El Nino events throughout the mid-20th century, centrally located anomalies ('El Nino-El Modoki') occurred more frequently in the century's latter decades. Shifts in tropical Pacific anomalies govern shifts in the Walker circulation, thereby scripting different outcomes for remotely affected regions [6]. Similar inconsistency is seen in the often negatively correlated relationship between El Nino and the Indian summer monsoon, with the El Modoki 'flavor' tending to be more consistently associated with suppression of Indian summer rains than its more easterly positioned earlier-century counterpart [7].

Simultaneously occurring with the mean-state's collapsed zonal asymmetry is a strengthened meridional component (Hadley Circulation) that re-directs atmospheric heat-flow away from the tropics. This atmospheric reorganization is communicated to planetary-scale subsurface ocean waves (Kelvin and Rossby). They, in turn, re-structure subsurface flow of heat and ocean mass. The result of El Nino's massive oceanic-atmospheric re-organization is a decrease in influx of ocean-heat and ocean-mass to the tropics. During a warm-ENSO event, net flow of heat is from the tropics poleward. In contrast, net heat-flow is equatorward during a La Nina. During a cool-ENSO event, warm surface water and the overlying Walker (zonal) Circulation shift far westward; Hadley (meridional atmospheric) circulation weakens; cold waters dominate the eastern basin, absorbing atmospheric heat and carrying it into the equatorial ocean subsurface. Warm subsurface waters from the subtropics flow toward the equator. Envisioned by some as a heat pump [8-10], ENSO collects heat in its neutral and cool-polarity phases and expels it poleward via ocean and atmosphere during its warm-polarity phase, with local, remote, short-term, and long-term impacts on weather and climate dynamics. Multidecadally, the frequency and intensity of events are modulated, with frequency of El Nino events (or La Nina events) being greater in some decades than in others [1, 11], with indications for heat re-distribution on decadal time scales.

The question of why easterlies occasionally subside or strengthen, and why the interannual (two-to-seven year) oscillation period is irregular, stirs debate. Source water from the subtropics [9] and/or velocity of subtropical-water transport [12-13] could play roles in subsurface water temperature feeding the cold tongue, altering the zonal SST gradient, thereby influencing easterly wind strength. In general, it is suggested that whatever the mechanism, it likely involves the decreased/increased temperature gradient between SSTs of the western warm pool and the subsurface waters feeding the cold tongue in the east, a condition whose causes could be local or global, external or internal, short-term or long.

Atlantic Nino and Benguela Nino:

Mean-state traits of the tropical Atlantic are similar to those in the Pacific: a coupled oceanic-atmospheric system with prevailing easterly winds, warm water pooled in the west, and a cold tongue of upwelled deep water in the east. As with the ENSO system, overlying SLP variations and interplay in the Atlantic between perturbation-generated subsurface planetary-scale waves – Kelvin and Rossby waves – result in growth, then demise, of a warm event. In the Atlantic, there are two types of warm events – the Atlantic Nino, associated with northeasterly winds, and the Benguela Nino, associated with southeasterly winds. With either type, when a warm tropical Atlantic event occurs, easterlies subside; sea-surface temperatures increase in the east. The zonal atmospheric circulation - the Walker Circulation - shifts eastward, weakens, and is supplanted by a strengthened meridional atmospheric component – the Hadley circulation. An intensified Hadley cell carries heat poleward and strengthens the Atlantic sector's subtropical jet. Due to the impact of a narrower basin-width on interacting subsurface waves, events in the Atlantic are weaker, shorter, and more frequent than in the Pacific. Despite this, end-results between an El Nino and an Atlantic event are similar: re-distribution of heat from the tropics poleward through oceanic and atmospheric circulation changes, with regional and remote impacts.

There are times when dynamics of the Atlantic shift in apparent response to a tropical Pacific-centered El Nino, suggesting inter-basin communication. Within several months of an El Nino, an Atlantic Nino sometimes follows – 1997/1998, for example. Yet, at other times, no relationship is apparent – 1982/1983 was such an instance. Reasons for the inconsistent response to similarly large El Nino events are unclear. Hypotheses include destructive interference [14], influence from a sometimes El Nino-modified feature – the Pacific North American (PNA) pattern [15], or an atmospheric bridge [16].

Indian Ocean Dipole (IOD) or Indian Nino:

Equatorial zonal interannual variability of a coupled oceanic-atmospheric system also exists in the Indian Ocean – the Indian Ocean Dipole (IOD [17-19]), but its character is not fully analogous to its Atlantic and Pacific counterparts. While the IOD is comparable in fundamental ways to tropical zonal variability in the Atlantic and Pacific Oceans, geography surrounding the Indian Ocean imposes significant differences – a discontinuous eastern boundary and the vast landmass of India to the north. The former allows direct oceanic communication with the Pacific basin to its east. The latter – a southern extension of the Indian subcontinent - blocks east-west oceanic flow within the Indian Ocean basin north of 25ºN. The imposing influence of this landmass scripts prevailing wind regimes - monsoonal flow: northeasterly in boreal winter to southwesterly in the boreal summer. These seasonally reversing winds within the Northern Indian Ocean tend to mask the equatorial zonal mode.

Equatorial Indian Ocean conditions differ from those in the monsoon-dominated regions northward. A weak zonal-wind component lingers over the equatorial Indian Ocean when monsoons are strong. Yet during transitions from one monsoonal flow to the other – i.e. during boreal spring and fall – equatorial westerlies intensify. The end result of these diverse wind patterns across the Indian Ocean is a mean equatorial state of warm water pooled in the eastern basin, below which the thermocline is deep and above which, the zonal convection of a Walker Cell is strengthened. These mean conditions – a reversed profile of the Pacific's - supply abundant moisture to Indonesian and little to East Africa.

Interannual fluctuations occur between amplified and weakened expressions of this equatorial mean state. First signs of a developing event emerge characteristically in the late boreal spring or early summer. If the seasonally weak equatorial westerly winds intensify, the equatorial mean state is amplified. A negative (cool) Indian Ocean Dipole (IOD) has potential to develop, delivering more moisture to Indonesia; less to East Africa. On the other hand, if the equatorial westerly winds fail, and are supplanted by equatorial anomalous easterlies and southeasterlies, a positive (warm) IOD may take hold. Zonal asymmetry collapses. The SST anomaly gradient reverses. Atmospheric convection shifts west. Strong rains pelt East Africa. Indonesia and Australia are unusually dry. An 'event' typically persists until the following winter.

Teleconnections reach beyond the surrounding region, some apparently decadally modulated or, in some cases, either enhanced or damped by interactions with teleconnections from ENSO or with processes in the tropical Atlantic [20]. Although the IOD is thought to be an independent intrinsic coupled mode of variability of the Indian Ocean [17-18], warm ENSO events often co-occur with positive IOD events and cool ENSO with negative IOD. ENSO has been posited as both being a potential trigger for [21-23], and potentially triggered by the IOD [24].

Meridional (north-south) Modes of Coupled Ocean-Atmosphere Variability in the Tropics:

Atlantic Meridional Mode (AMM) and Pacific Meridional Mode (PMM):

Meridional modes of the tropics involve latitudinal migration of the ITCZ in relatively swift response to winds and associated SST anomalies. Meridional modes, which are dominantly interannual, yet multi-decadally modulated by oceanic influence [25], have been identified in both the Atlantic (Atlantic Meridional Mode (AMM)) and the Pacific (Pacific Meridional Mode (PMM)) [26]. They are variable coupled oceanic-atmospheric phenomena in the tropics that are excited by boreal winter extratropical atmospheric variability (see section: annular modes) over their northern basins. Wind-induced latent heat flux leads to a summer-centered coupled response in tropical SST anomalies. A meridional SST gradient across the mean latitude of the ITCZ develops. Strongly sensitive to small meridional SST gradients, the ITCZ migrates toward the warmer hemisphere, accompanied by cross-equatorial surface winds flowing in the direction of that hemisphere. Remotely forced by extratropical atmospheric circulation in their respective basins, meridional modes impact regional temperature, precipitation, and hurricane [25-27] regimes, with potential impact extending to zonal tropical modes, e.g. ENSO [28].

Circulation Patterns of the Extratropics

Geographically Governed Dynamics:

The Coriolis Effect is non-zero away from the equator and throughout the extratropics. Its parameter, which is same-signed throughout a given hemisphere, increases with latitude. In part because of Coriolis-related peculiarities, travel of Kelvin waves in the extratropics is restricted to poleward (equatorward) flow within a narrow zone along western (eastern) coastlines of continents or cyclonically around a closed boundary. Kelvin waves cannot travel in the open ocean along lines of latitude in the extratropics. Only westward-propagating Rossby waves can, with slower velocities at higher latitudes. Consequence of these dynamics is two-fold: 1) communication of overlying atmospheric changes to the subsurface ocean is propagated only toward the west; and 2) time required for the Rossby waves to travel cross-basin takes years to decades in the mid-to-high latitudes. This renders adjustment of the subsurface ocean to overlying large-scale wind field slow in the extra-tropics. Extratropical ocean 'dynamic memory' is long; persistence of a climate signal at mid-to-high latitudes is thereby extended.

In addition to the extratropical traits responsible for modifying climate signals and their transmission, equator-to-pole atmospheric transport of heat is similarly amended. It is broken into three large-scale convection cells – this due to the Coriolis Effect. Related rising, sinking, diverging, and converging basin-scale air flow within these cells establish semi-permanent surface high and low-pressure regions and associated belts of prevailing winds, thus governing fundamental aspects of regional circulation patterns.

Dominantly Atmospheric Circulations:

Annular Modes and Related Extratropical Circulation:

Atmospheric variability at mid-to-high latitudes in each hemisphere is strongly influenced by annular modes: the Northern Annular Mode (**NAM**) in the Northern Hemisphere and the Southern Annular Mode (**SAM**) in the Southern Hemisphere [29-31]. Shifts of atmospheric mass between high latitudes poleward of ~ 60º and a ring, or annulus, around the mid-latitudes (~45º) characterize these patterns with no preferred timescale of variability, manifesting as large-scale variability of angular momentum in their respective hemispheres. Lower-than-normal sea-level pressures over the Polar Regions, higher-than-normal pressures in mid-latitudes, and positive westerly-wind anomalies at about 55º to 60º characterize the high-index (positive) polarity. The reverse is true for the low-index (negative) polarity.

During the boreal winter for the NAM and the austral spring for the SAM, these tropospheric patterns merge with stratospheric dynamics. Although often used interchangeably with NAM and SAM, the Arctic Oscillation (AO) and Antarctic Oscillation (AAO), respectively, are terms that apply to the tropospheric expression of the seasonal stratospheric-tropospheric coupling. This stratospheric-tropospheric coupling involves a polar vortex of westerly winds that extends seasonally from the stratosphere to the surface, thereby enabling stratospheric perturbations to influence activity in the troposphere, from the poles to the equator. In turn, when the two atmospheric levels are coupled, tropospheric processes impact the stratosphere. This vertical two-way coupling imparts a low-frequency component to the annular mode's time scale of variability. Associated with this lower frequency component is a recently proposed relationship between ocean-heat flux from western-boundary currents (WBCs) and their extensions (WBCEs), and sudden-stratospheric-warmings (SSWs) [32]; the latter, in turn, force the troposphere. Both phenomena – WBCE ocean-heat flux and SSWs - exhibit decadal to multi-decadal variability. The WBCE-related ocean-heat flux out of the ocean is positively correlated to lateral advection of upper-ocean heat to WBCs. This lateral advection of ocean heat, itself, co-varies at a one-year lag with westerly wind strength associated with a multi-decadal component of the NAM [33-34].

The North Atlantic Oscillation (**NAO**: [35]) is a North Atlantic-centered manifestation of the NAM. Both NAM and NAO characterize similar temporal and spatial leading modes of Northern Hemispheric variability. While NAM is hemispheric, NAO is confined to the North Atlantic region. Atmospheric-mass re-distribution between subpolar and subtropical latitudes manifests as changes in the normalized SLP anomaly-difference between the North Atlantic atmospheric centers-of-action (COA): the Icelandic Low and the Azores High. A positive SLP anomaly-difference represents the high-index, or positive polarity, of NAO and a negative SLPA-difference indicates the negative polarity, or low-index, of the NAO. Low-frequency modulation of interannual-to-interdecadal variability is apparent. A characteristic tri-polar pattern of wind-generated SST anomalies within the North Atlantic is generated by higher-frequency fluctuations of the NAO [36, 37]; while a more uniform, basin-wide SST anomaly pattern, likely involving large-scale ocean dynamics [37-40], distinguishes the low-frequency component, with significant teleconnection impact. [39-40].

The NAM also extends its influence into the North Pacific; although not as conspicuously as in the North Atlantic. In fact, SLP fluctuations over the North Pacific and North Atlantic exhibit no significant correlation – an observation seemingly inconsistent with the paradigm of a hemispheric NAM. But reconciliation might be found in an argument that has been made for the indirect correlation between the Atlantic and Pacific ocean basins. While the Atlantic and Pacific centers may not correlate positively with one another, both correlate negatively with the Arctic center-of-action. Through that indirect association, it is proposed that observed SLP behavior in the North Pacific is compatible with the hemispheric paradigm of polar-subtropical see-saw exchange of atmospheric mass [41]. Obscuring that otherwise potentially obvious NAM- North Pacific relationship may be the co-existence of another strong mode of climate variability. That mode may be found in the Pacific-North American (**PNA**: [15]) pattern.

PNA is a prominent upper-atmospheric mode of low-frequency variability involving anomalous atmospheric pressure. Located at mid-tropospheric heights over the extratropics, influence on its behavior is not confined to the mid-to-high latitudes. Strongly associated with the Aleutian Low in the North Pacific, the PNA also is modified by tropical processes. Associated with jet-stream tracks, modified by underlying topography and constraints of vorticity conservation, the PNA is linked to the flow of storm systems (wave trains) across the Northern Hemisphere. Intensity and location of weather systems, particularly those affecting North America in the boreal winter, are most impacted by the PNA. Liu et al. [42] attempted to tease apart behavior of the AL and AO (NAM). Teleconnected influence of each individual center-of-action differed from the teleconnections resulting from the combination of their influences. This observation is consistent with the idea of co-existence of patterns and the consequent possible masking of the NAM signature in the North Pacific.

While the PNA is strongly associated with the North Pacific via the Aleutian Low, its scope is hemispheric. More regional in nature is the North Pacific's analogue to the NAO, namely, the North Pacific Oscillation (**NPO**: [43]). In the North Pacific area, including the most northwestern sector of North America and the eastern coastal areas of Eurasia, the NPO is the most important teleconnection mode. As with the NAO, NPO can be described as a north-south seesaw of atmospheric mass between subpolar and subtropical regions, with the normalized SLP anomaly difference between the regions measuring the polarity and strength of the pattern's interannual-to-interdecadal fluctuations.

Originally described in 1932, NPO was said to result primarily from geographical shifts of the mean position of the AL [43]. Indeed, it is observed that AL migrates west and north when weak (increase SLP) and to the east and south when strong (decrease SLP), the most extreme shifts occurring on multidecadal timescales [44]. Spatial scale of NPO-influence fluctuates similarly [45]. During decadal-plus intervals, the footprint of the NPO appears confined to the North Pacific basin. The AL is weak and skewed west of its mean position. During these 'regional' time spans, ENSO plays a strong role in the NPO behavior, with pronounced influence on AL. In contrast, during 'hemispheric' intervals, the reach of NPO influence spreads far beyond the Pacific basin. The AL is strong and shifted southeast of its mean position. ENSO forcing on AL is minimal. Instead, extratropical forcing of the low-pressure system is dominant. With an intensified AL, the hemispheric PNA pattern is more pronounced.

NAO, too, shows indications of vacillating on low-frequency timescales between more regionally confined intervals and more hemispherically spanning ones [46]. Not unexpectedly, details of NAO/NPO high-frequency variability and of associated teleconnected impacts related to precipitation, surface-temperature, ocean-gyre activity, and sea-ice patterns, vary accordingly at the low-frequency tempo. Geographical shifts of the Arctic, subpolar, and subtropical atmospheric COAs on decadal-plus timescales can explain much of the observed behavior. Evidence suggests indirect oceanic influence on atmospheric COAs [47-51]. Low-frequency solar variability, too, may play a role [44, 52].

The SAM, many of its features analogous to the NAM, overlies relatively simple geography. Absence of landmass in the mid-latitudes around 60ºS conveys unparalleled strength to the SAM-associated mid-high-latitude westerly winds. These winds drive the Antarctic Circumpolar Current (ACC) in the Southern Ocean – a globally girdling ocean current that connects all three ocean basins and is fundamental to global intercommunication of climate signals. Through upper-atmospheric communication (via the Pacific South American Pattern (PSA)), this southernmost atmospheric pattern, the SAM, whose influence strongly impacts Antarctic temperatures, sea-ice dynamics, and associated deep-water-formation, appears to be linked to tropical dynamics, particularly ENSO in the Pacific sector [53-55].

Dominantly Oceanic Circulations:

The atmospheric circulations described above are dominated by stochastic, high-frequency behavior. In contrast, dominantly oceanic circulations operate at low-frequency timescales. Water's high heat capacity, in conjunction with long time spans required for both extratropical subsurface Rossby-wave travel and vertical mixing, contribute to the ponderous pace of oceanic processes. At these low-frequency time scales, the ocean can exert direct and indirect influence on the atmosphere, thereby adding persistence to overlying atmospheric circulation. This influence is reflected in a low-frequency modulation of that atmospheric circulation's frequency and intensity at interdecadal to multidecadal time scales. In addition, the ocean modifies hemispheric and global teleconnections associated with the atmospheric patterns.

Atlantic Multidecadal Oscillation (AMO):

Ocean-heat transport is northward at all latitudes of the Atlantic Ocean. This cross-equatorial ocean-heat transport is unique to the Atlantic, generating a dipole of SST anomalies between the North and South Atlantic basins. This SST distribution is not static. Proxy and instrumental records support the existence of an intrinsic low-frequency mode of variability, characterized by a multi-decadally (50-to-80 year) varying monopolar pattern of sea-surface-temperature anomalies averaged across 0º to 60ºN and 75º to 7.5ºW. It is known as the Atlantic Multidecadal Oscillation (AMO: [56-60]). Climate impacts of the AMO-related SST anomalies are regional – North America and Eurasia are particularly impacted. AMO's reach can be hemispheric, as well, extending to the North Pacific [57-58, 61-62].

Recent studies introduce the notion that AMO operates at a second timescale of variability, a 20-to-30-year quasi-regular period [63-66]. Both the higher-frequency mode and the more well-known low-frequency mode elude full description; although both appear related to fluctuations in strength of the Atlantic sector's Meridional Overturning Circulation (AMOC) [65-69], as is suggested by model studies incorporating an interactive ocean [67, 70] and by variations in instrumental [71], and in proxy [72] records of the interhemispheric SST dipole pattern. Mechanistic details for the AMO-AMOC connection remain under debate [67, 73-76].

Westward propagation of SST anomalies is associated with the hypothesized higher-frequency AMO mode [75]. Also associated are sea-level variations [64] along the North American and European coast lines. One idea is that the signal is generated from an internal ocean mode within mid-to-high latitudes of the North Atlantic, perhaps excited by low-frequency atmospheric noise [75]. From the North Atlantic, the thermal anomalies propagate into the Arctic [63].

In contrast, studies support a strong active role for Arctic involvement in the lower-frequency mode [66, 75-76]. This more slowly paced pattern might involve a salinity oscillation between the Arctic and high latitudes of the North Atlantic [66, 76-77]. According to modeled results, this proposed oscillation is associated with saline Rossby modes in the Arctic, which perhaps are excited by variability in processes that influence freshwater balance in the region (e.g. Atlantic inflow, sea-ice changes, or variability in river runoff) [66]. The approximate 60-year rhythm found in 20th-century AMO SST anomalies and Arctic/Atlantic salinity oscillations can be found in numerous patterns: the 1) Atlantic core-water temperature [78-79]; 2) Arctic sea-ice-extent in the Western Eurasian Arctic Shelf Seas (Greenland, Barents, Kara Seas) [80-82]; 3) basin-scale meridional-temperature-gradient (MTG [83]), 4) basin-scale westerly wind strength [80-84], and 5) surface temperature over Eurasia [83].

Potentially related to the salinity oscillations and the above-mentioned set of processes is evidence for an approximate 60-year variability in the geographical positioning of the axis between sea-level-pressure (SLP) centers-of-action (i.e. Icelandic Low and Azores High) overlying the AMO-related SST anomaly pattern. This axis realignment, with its low-frequency component projected onto the NAO, is hypothesized to influence the North Atlantic's high-latitude salinity concentration, and by extension, the multidecadal character of the AMOC strength and AMO polarity [47, 58].

Pacific Decadal Oscillation (PDO):

The Pacific Decadal Oscillation (PDO (Fig. 2)) is an oceanic circulation pattern characterized by a boreal-winter, basin-wide, quasi-periodically varying SST pattern north of 20ºN in the North Pacific [85-86]: A positive-PDO hosts a pattern of cool SST anomalies in the vast central/western sector of the basin and warm El Nino-like anomalies in the tropical sector and along the west coast of North America. A negative-PDO pattern is reversed. A related pattern is the Interdecadal Pacific Oscillation (IPO [87]), which includes both the North and South Pacific Oceans and of which the PDO is likely a part. Both the PDO and IPO have been shown to display two dominant quasi-periodicities, one of 15 to 30 years and another of 50 to 70 years (See Fig. 1). Both are intimately coupled with the overlying atmospheric circulation, which involve the Aleutian Low (AL) and the Hawaiian (or Pacific) High (HH).

Longitudinal/latitudinal migrations of these COAs co-vary with phases of the PDO, the lower frequency shifts being more extreme. With a positive-PDO, an intensified AL shifts east and south; the strengthened HH moves west and south [52]. Movements are opposite for a negative-PDO. Re-positioned COAs can yield unexpected results. Consider the behavior of El Nino-associated phenomena. They can differ according to AL position, and by extension, PDO phase. An example can be found in fluctuating sea-ice growth in the Bering Sea. When PDO is in its negative phase, El Nino events correlate with decreased ice extent. In contrast, when PDO is in its positive phase, Bering Sea ice-extent increases during El Nino events. This is due to re-positioning of AL and the consequent placement of accompanying winds [88].

Re-positioned COAs can also modify the extent and the consequences of PDO-related teleconnections, many of which involve ENSO. The multidecadal component of PDO modulates the frequency and intensity of ENSO. During a positive-PDO, warm-ENSO events increase in both intensity and frequency. During a negative-PDO, cool-ENSO events are favored [10, 1].

With PDO-modulation of ENSO due to COA migrations, remote teleconnections of ENSO can build in impact due to cumulative effects. Some PDO-modulated ENSO teleconnections impact multi-decadal variations in the freshwater balance of the North Atlantic. Impacts occur on differing time scales. They include: the 1) alteration of hemispheric-scale precipitation patterns (increased precipitation in tropical Atlantic with La Nina [89] and decreased with El Nino [90]); 2) the basin-scale flow of relatively fresh water out of the Pacific (increased fresh Pacific water to Arctic [88, 91] and decreased to Indian Ocean during El Nino); and 3) a possible increase in occurrence of positive IOD events with El Nino. With them comes a decadal-scale-lagged suppression of salt delivery to Atlantic from Indian Ocean (Agulhas Leakage) [92, 93]. The opposite occurs with La Nina.

PDO low-frequency control of ENSO teleconnections also influences sea-ice-formation regions off the Antarctic coasts (more in Weddell Sea (Atlantic sector) during El Nino and more in Ross Sea (Pacific sector) during La Nina), with delayed potential impact on strength of the AMOC. Delivery of upper-ocean heat by the North Pacific subtropical gyre to the western-boundary current and its extension is similarly paced [94].

From the examples outlined above, it can be seen that Atlantic and Pacific multidecadally varying ocean modes, each through indirectly teleconnected influences, appear to modify the other at decadal-plus timescales, influencing their temporal relationships. That temporal relationship during the 20^{th} century shows the 50-to-80-year mode of PDO has occurred in quadrature with that of the AMO [1, 95]. In other words, their phases show an approximate offset by a quarter-of-a-period (Figure 1). Their relationship is correlated with drought patterns in the contiguous United States. Observation indicates that drought has been most extreme during positive phases of the AMO; while the accompanying phase – positive or negative - of PDO has scripted the distribution of drought-impacted regions (**Figure 3**) [96]. The record of Colorado River basin flow over the last century appears consistent with this observation [97]. In addition, 500-year-long proxy records reflect a similar cadence in the Colorado River basin stream-flow, suggesting a strong and persistent natural component for this observation [98].

Antarctic Circumpolar Wave (ACW):

Identified in 1996 [99], the Antarctic Circumpolar Wave (ACW), a product of polar-subtropical temperature contrasts, consists of SLP and SST anomalies in a wave number two structure that propagate in the Southern Ocean around Antarctica. Its interannual quasi-periodicity of four-to-five years is modulated on decadal, interdecadal, and multidecadal time scales. The ACW propagates westward in the vicinity of the eastward flowing, SAM-related, hemispherically encircling Antarctic Circumpolar Current (ACC). Influenced by the fast eastwardly flowing ACC, net motion of the westward ('upstream') - propagating ACW is eastward, yet at a slower rate than that of the ACC. The ACW makes a full trip around the globe at southern latitudes (~65ºS to 30ºS) in eight to nine years. Weather patterns in southern regions of Australia, South America, and Africa are influenced by the ACW.

Influence of and on the ACW extends to the tropics, where it appears to engage in two-way communication with ENSO. The atmosphere transmits tropical information to high southern latitudes, imprinting an ENSO footprint upon the ACW. Piggy-backed upon the ACW, the original ENSO footprint is slowly modified as it circumnavigates the Antarctic continent. ACW, through various bifurcations equatorward, returns the amended ENSO signature to the tropics, the return routing of the signal is guided by phase of PDO [100, 101].

Conclusion:

Earth's regional climate patterns work on a variety of timescales to globally re-distribute heat from where there is an excess to where there is a deficit. All Earth system components - ocean, atmosphere, land, and sea-ice - participate in tailoring characteristics of individual circulation patterns. Impact of individual patterns on weather and climate on local, regional, hemispheric, and global spatial scales is modified according to interaction among patterns and according to timescale. Complexity of local and regional detail tends to yield to hemispheric order on the long-term. This may be attributable, at least in part, to synchronized network behavior of climate indices at low-frequency time scales, as recently hypothesized by [1; 81-82]. Interaction of regional circulation patterns, especially at multidecadal tempos, is often accompanied by characteristic phenomena. Interconnectedness of Earth's climate network is profound, the foundation of the planet's heat engine.

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Links to graphics and additional material for circulation patterns:

El Nino: http://www.nature.com/scitable/content/ne0000/ne0000/ne0000/ne0000/13286620/stevens_figure7_climate_ksm.jpg

Atlantic Nino and Atlantic Meridional Mode: http://booksite.academicpress.com/DPO/gallery/chs15/s15_01ad_full.jpg

Benguela-type Atlantic Nino: http://folk.uib.no/ngftg/atlant2.html

IOD: http://2.bp.blogspot.com/_6VXBV3uWTc4/Swt-oxXl7WI/AAAAAAAAABo/YTnGLt8bO9w/s1600/Indian_Ocean_Dipole.png

Earth's global circulation: http://www.atmosphere.mpg.de/enid/2__Circulation_Systems/-_Global_Circulation_18z.html

Northern Annular Mode: http://www2.ucar.edu/sites/default/files/ucar_magazine/currents/2010/arctic_oscillation.jpg

NAO: http://www.ldeo.columbia.edu/res/pi/NAO/ and

http://ossfoundation.us/projects/environment/global-warming/north-atlantic-oscillation-nao/image/

Atlantic Multidecadal Oscillation: http://ossfoundation.us/projects/environment/global-warming/atlantic-multidecadal-oscillation-amo/image

PDO: http://jisao.washington.edu/pdo/

ACW: http://openi.nlm.nih.gov/imgs/rescaled512/1074811_pbio.0030127.g003.png

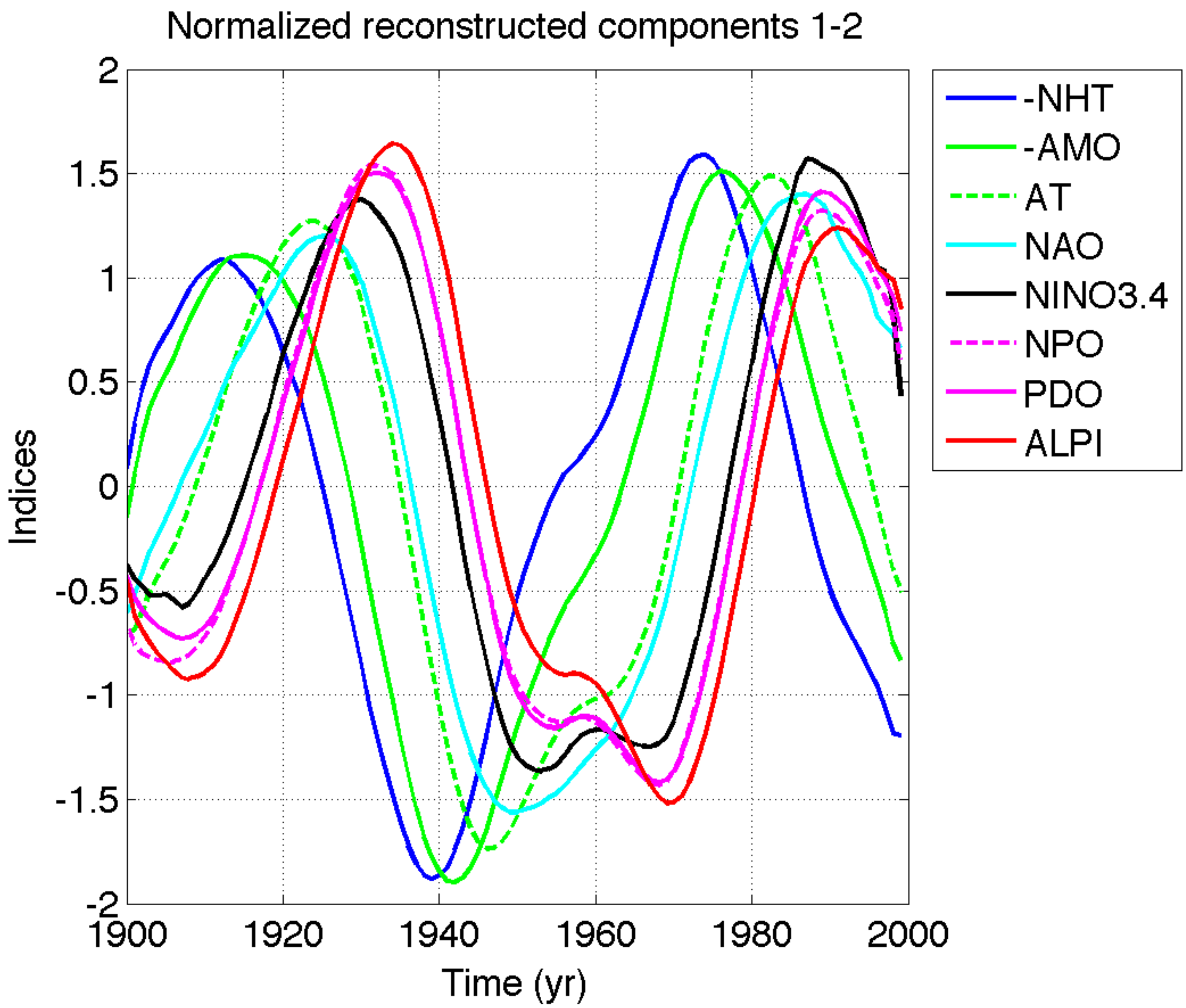
Figure Captions:

Figure 1: A low-frequency signal propagating through a network of synchronized climate indices is reflected in this graph. The propagation pattern is termed the "stadium-wave" signal by its original authors – a reference to its communication through a sequence of indices. The plot shows normalized reconstructed components (RC) of an eight-index climate network. Indices include anomalies of the averaged Northern Hemisphere surface Temperature (NHT), the Atlantic Multidecadal Oscillation (AMO), Atmospheric-Mass Transfer index (AT: refers to zonal component of basin-scale wind

direction), the North Atlantic Oscillation (NAO), NINO 3.4, an index for El Nino, the North Pacific Oscillation (NPO), Pacific Decadal Oscillation (PDO), and the Aleutian-Low Index (AL). The RC time series shown have been normalized to have a unit variance. The indices are synchronized at non-zero lags (except for NPO and PDO, whose rescaled RCs are virtually identical). Note the reconstructions of NHT and AMO are of negative polarity. Reproduced with permission: Wyatt et al. 2012 [1].

Figure 2: Anomalous climate conditions associated with warm (positive) and cold (negative) phases of the Pacific Decadal Oscillation (PDO). Values are shown for degrees Celsius for sea-surface temperature (SST), millibars for sea-level pressure (SLP), and direction and intensity of surface wind stress (arrow width proportional to wind strength). The longest wind vectors represent pseudostress of 10m^2/s^2. (Anomalies would need to be multiplied by associated index value (not given here) in range between -3 to +3. Reproduced with permission: Steven Hare; adapted and updated from Mantua et al. 1997 [85].

Figure 3: Drought frequency (in percent of years) in the contiguous United States for warm and cool (positive and negative) low-frequency regimes of the Atlantic Multidecadal Oscillation (AMO) and the Pacific Decadal Oscillation (PDO). A) +PDO and –AMO; B) –PDO and –AMO; C) +PDO and +AMO; and D) –PDO and +AMO. Note that drought is most widespread when AMO is in its warm (positive) state (figs. 3C, 3D). The distribution of drought is modulated with phase of PDO (more in northwest and north central and southeast with positive PDO and positive AMO; more southwest and Midwest with negative PDO and positive AMO). The "stadium wave" in figure 1 gives insight into when the various AMO and PDO phases coincide. Reproduced with permission: Greg McCabe et al. 2004 [96].



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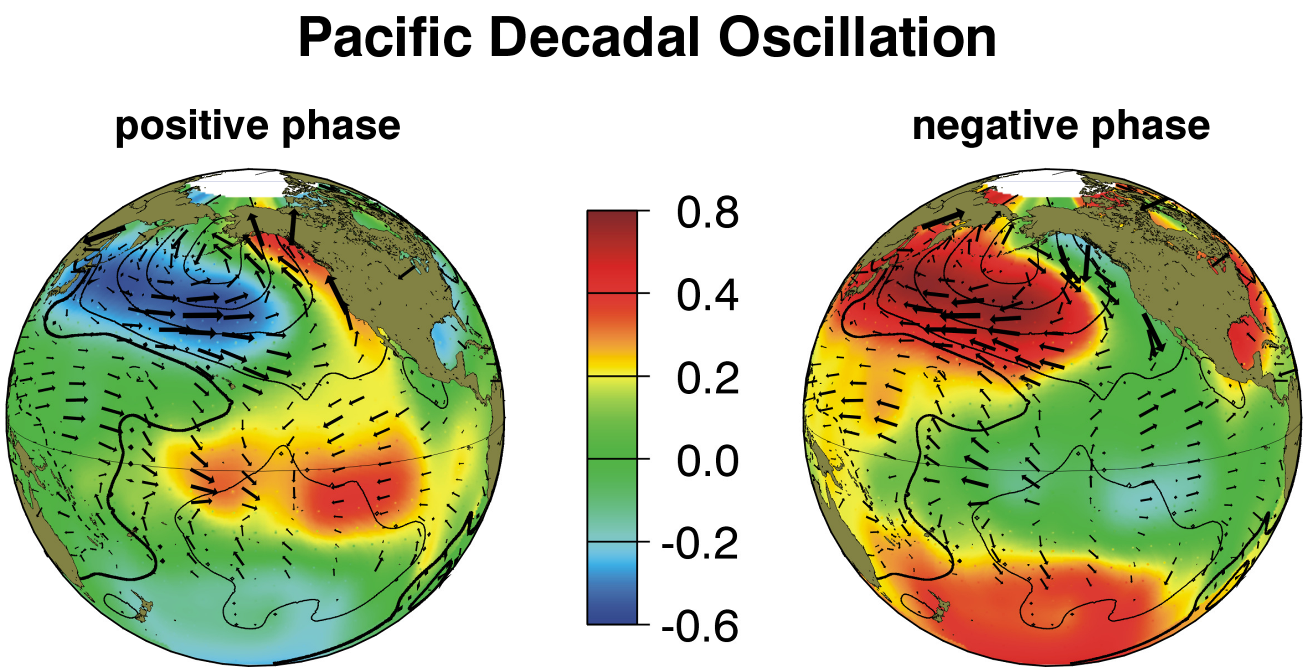

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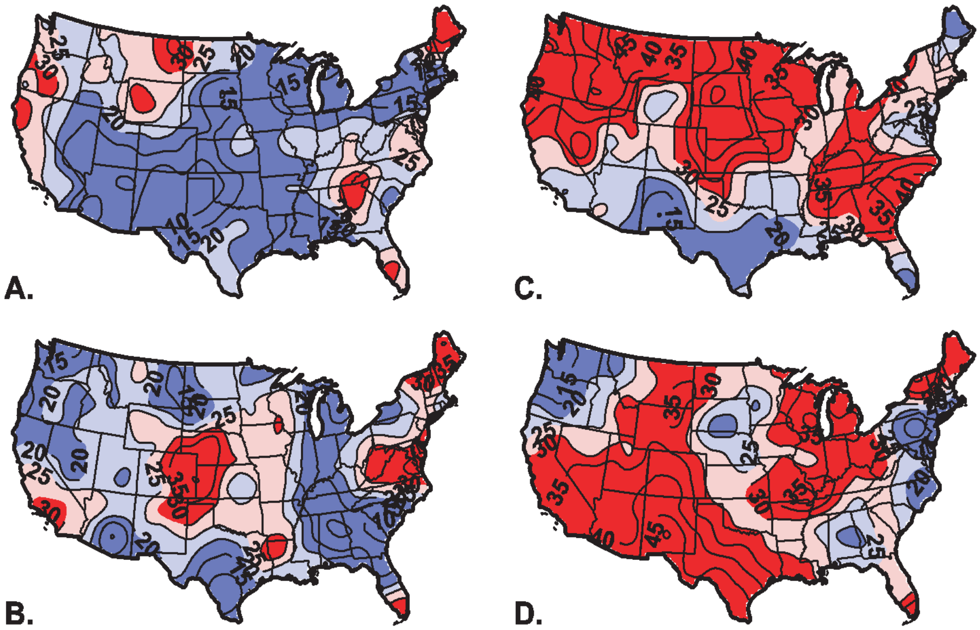

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