Patterns of Global Climate Variability: Are the Tropics "Control Central"?

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ABSTRACT:

Fluctuations in mean global climate from interannual to decadal and possibly millennial timescales largely can be attributed to two dominant modes of variability – ENSO and the annular modes. Signals sent between the two modes, directly and through intermediary patterns, convey a network of opposing feedbacks that may contribute to stability of the global climate system, absent extreme cases of external perturbation.

This paper endeavors to illuminate the roles played by these dominant modes of variability and to highlight contributions made by ancillary patterns that serve to enhance the global network of communication. The tropical Pacific is home not only to one of the dominant modes of variability – ENSO – but home also to patterns that initiate and facilitate communications among disparate regions, including latitudes at which annular modes govern. The ability to teleconnect to remote regions and thereby engage a sequence of feedbacks that have the potential to stabilize global climate suggests that the dynamics of the tropical Pacific might suitably be framed in the context of "control central" (*PPT Slide 1*).

INTRODUCTION:

Distinct regional patterns of winds, ocean currents, sea-level pressure, precipitation distributions, and temperature fluctuations emerge from a background of global climate noise. The list of patterns is long: TAV, Atlantic-Nino, SAIM, PDO, IPO, NPO, SO, AO, AAO, NAM, SAM, PNA, PSA, GEW, ACW, NAO, IOD, QBO, MJO, and ENSO (See glossary, p 30). Likely I have missed a few. All appear to be quasi-cyclic with that quasi-cyclic periodicity, itself, changing on quasi-cyclic time frames. Variability on scales of days to multiple decades can be identified from this collection of oscillatory behavior. In addition, some variability is internal – something inherent in any random system. Some variability is extraterrestrial – with incoming solar variability on a variety of timescales being the most relevant. There are the added complications of geomagnetic variability, the slowly fluctuating variability of orbital parameters, and the slower-yet variability of Earth's tectonic rearrangements and cycles. All work together on a mind-dizzying level to generate patterns of coupled interactions among the various Earth systems – the lithosphere, cryosphere, hydrosphere, biosphere, and atmosphere – to ultimately create what we know as "climate". Teasing out the dominant players of this scenario that operate on timescales relevant to human perception is the challenge lying before climate researchers today. *(Slide 4)*

Most climate variability appears traceable to two dominant modes – ENSO and the annular modes[1-3]. Each mode is of regional origin yet of global influence. ENSO - interannually oscillating phases manifested most markedly as changes in SSTs in the eastern equatorial Pacific - reigns in the tropical Pacific. Through atmospheric and oceanic media, it transmits its influence both zonally and meridionally, touching every corner of the globe to some degree. Annular modes, responsive to perturbations of tropical origin, as well as to solar fluctuations, volcanics, aerosols, greenhouse gases, sea-ice extent, and SSTs, are pronounced seasonal tropospheric expressions of the stratospheric polar vortex in the high latitudes, yet extend their fingerprints to the tropics and beyond. The remaining above-mentioned oscillatory patterns, born of tectonics, ocean/atmosphere dynamics, and thermal gradients, play critical to ancillary roles as linkages and amplifiers to these two dominant modes.

The tropical Pacific, home of ENSO, may well be "control central" of global change[4]. Not only does the tropical Pacific maintain thermal stability of its low-latitude home-base through highly changeable atmospheric/oceanic dynamics[5-9], but those changes that convey stability to the tropics send signals to high latitudes and upper levels of the atmosphere, forcing changes in those regions, some of which are ultimately communicated back to "control central" *(Slide 5)*.

This paper will examine the dominant and ancillary modes of climate variability. Discussion begins with "control central", the tropical Pacific **-** a little history and an overview of ENSO basics, rounded out with a glance at the paleo-history and more recent history of ENSO behavior. Following is a relatively in-depth examination of the communication - zonally, meridionally, and vertically - of signals from the tropical Pacific to remote regions, revealing links to the numerous ancillary oscillatory patterns. Annular modes are considered next. Signals transmitted to, or deflected by, these modes reveal a relationship between high latitude variability and tropical dynamics. Fluctuations of the thermohaline circulation (THC), as orchestrated by ENSO and annular-mode dynamics, round out discussion of climate variability. Numerous questions evolve from the complexities described, suggesting directions for future research.

THE TROPICAL PACIFIC: SO & ENSO:

A Little History:

It was only a little more than a century ago, that atmospheric pressure fluctuations in Sydney, Australia were recognized to be out-of-phase with those in Buenos Aires, Argentina[10]. Within a few years came the realization that these fluctuations oscillated fairly regularly – about every 3.8 years[11]. Furthermore, it was noted that this as-yet unnamed oscillation was almost global in extent[11-12].

"When the pressure is high in the Pacific Ocean, it tends to be low in the Indian Ocean from Africa to Australia," observed the Director-General of Observatories in India, Sir Gilbert Walker. He based his observations on pressure readings in Jakarta, Darwin, and Cairo compared to those in Santiago, Honolulu, and Manila. He identified one such pressure fluctuation - the pressure fluctuation between the Indian Ocean and the eastern tropical Pacific - as the Southern Oscillation (SO). Gilbert also took note of temperatures in Madras, rainfall patterns in India and Chile, and the fluctuations in the Nile floods. He was motivated to understand the occasional, but disastrous failures of the Indian monsoons. In 1923 and 1937 he published work that showed the SO to be correlated strongly with precipitation and wind patterns over the tropical Pacific and Indian Oceans, as well as with temperature fluctuations in southeastern Africa, southwestern Canada, and southeastern U.S.[13-14]. Despite these observations, short and incomplete records handicapped further evaluation. Skepticism by Gilbert's peers dampened general interest in the topic. At the time, only the atmospheric component was recognized; no one realized that the ocean may play a part. It was not until near the end of Gilbert's life that the ocean was identified as playing a role linked to the fluctuations of the SO, or more accurately, the SOI – the Southern Oscillation Index, the difference between the pressure centers. When that difference is higher-than-normal, the SOI is positive; when it is lower-than-normal, the SOI is negative.

Despite the hiatus of collective interest in the phenomenon, focus was not diverted for long. Using observations of Walker and others, Dutch meteorologist Berlage, in the 1950s, generated a map that illuminated the SO as a barometric record of the exchange of atmospheric mass along the entire global tropical circumference. An impressive degree of organization within the SO is apparent from all the records. Its apparent periodicity of four years falls short of a predictable pattern, as the period is irregular, from two years to ten. Modern-day wavelet analysis shows its peak to be between three and eight years, and this peak has shifted considerably within the hundred-plus years of its documentation[17].

Anomalous atmospheric and oceanographic conditions in 1957/58 further fueled interest in the tropical Pacific. That year, prevailing easterlies weakened and SSTs in the East Equatorial Pacific (EEP) warmed. Professor Jacob Bjerknes, of UCLA, suggested that this collection of anomalous conditions represented an event that was not unique, but rather was an interannual event, now known as an El Nino event. A decade later, *Bjerknes '66, '69* [15-16] put forth a hypothesis that related the oceanographic and meteorological variations that create alternating conditions along the tropical Pacific. His goal was to establish a link among changes in observed equatorial SSTs, zonal wind strength and direction; oceanic upwelling in the EEP; and surface pressure anomalies within the SO. He posited that a zonal atmospheric circulation, which he named the Walker Circulation, derived its energy from zonally asymmetric SSTs (where west-to-east SST differences along the equator in the Pacific are on the order of 4° to 10°C) and is part of the larger SO. He noted that fluctuations in the Walker Circulation strongly influenced the SO. In turn, changes in this collection of circulations affect the meridional Hadley circulation, which affects latitudinal distribution of tropospheric heating.

Bjerknes recognized that the peculiarly strong zonal asymmetry that existed in the equatorial Pacific was maintained by a coupled positive feedback. Strong easterlies, born of strong SST asymmetry, further enhanced asymmetry of SSTs, which further strengthened the easterlies. He noted that when easterlies weakened and SSTs increased in the eastern Pacific, this too was maintained by a coupled positive feedback: weaker easterlies promote zonal symmetry of SSTs and zonal symmetry of SSTs further weakens the easterlies. His insight illustrated that interactions between the ocean and atmosphere in the tropical Pacific constitute a circular argument. A slight change in either the atmospheric or oceanic component would cause a change in the other. This was revolutionary. With this new paradigm, the evolutionary path of a cold or warm event became clearer. EEP SSTs could now be explained in terms of winds, and thus related to pressure gradients. A "warm event" was the El Nino phase of the SO. This epiphany pulled El Nino and the SO into a single mode of the ocean-atmosphere system – ENSO (El Nino Southern Oscillation). While a huge advance in thinking, Bjerknes' insight still fell short of explaining what caused the system to shift from one phase to another. A mechanism for this alternating nature remained elusive[16-17].

Over the intervening thirty-five years, observation of and research into the ENSO system has brought more insight and more questions. Georges Philander, in his book, *El Nino, La Nina, and the Southern Oscillation*[18], points out that the ENSO system is best left to ambiguous and qualitative description, much akin to the term "winter". The term "winter" is useful in that it describes a broad set of conditions with which we are all familiar; yet it is clear to us all that each winter is distinct in character, understood best in retrospect, far from a predictable quantity. So it is with the ENSO system.

ENSO Basics:

ENSO is a natural zonal mode of the coupled atmospheric-oceanic system in the tropical Pacific, characterized by fluctuations between two end-member phases: La Nina, the cold phase, and El

Nino, the warm phase. The cold phase is distinguished by strong zonal asymmetry of SST distribution, thermocline depth, sea-surface height (SSH), and sea-level pressure (SLP). Strong easterly winds are both a result of and cause of this asymmetry in parameters. The warm phase hosts weakened easterlies and anomalous westerlies. SSTs, SLP, thermocline level, and SSH are more zonally uniform. Both phases are stable with respect to the conditions that accompany them. From this observation, it is intuitive to assume that the tropical Pacific would become phase-locked into one regime or another until the hosting conditions changed. The system does not become phase-locked. It is therefore apparent that the hosting conditions must change. How is not entirely clear.

The system oscillates on an irregular interannual basis within a spectrum of conditions book-ended by these cool and warm end members, El Ninos being slightly more dominant in occurrence than La Ninas. The system may be self-sustaining, but more likely it is a system prone to imbalance, allowing it to become vulnerable to a trigger that moves the system from the cool phase to the warm phase[18]. The timing of that "trigger" is important. Feedbacks amplify the event; internal dynamics likely help bring the system back to the cool phase. This progression from one phase to the other and back has been described loosely in terms of "sloshing".

"Parts" of the system include the zonal atmospheric pattern - the Walker Circulation; the meridional atmospheric pattern - the Hadley Circulation; the prevailing easterly Trade Winds; the ITCZ; the SPCZ; the Western Warm Pool (WWP); and the equatorial East Pacific cold tongue. These components are coupled to the broad-scale atmospheric pressure system - the SO. In addition to these, there also exists a complex collection of surface and sub-surface equatorial and off-equatorial currents and countercurrents; a shallow, meridionally and zonally non-level thermocline, fed by waters from subtropical gyres on either side of the equator, prone to perturbation by westerly wind anomalies and host to perturbation-generated gravity ocean long waves – Kelvin and Rossby waves; and a region hosting both no Coriolis and opposing Coriolis forces. This inventory of features, situated within a region defined as lying between 30°N and 30°S, is choreographed in such a way as to render this tropical region "control-central" of numerous global climate patterns. Our current knowledge of the ENSO phenomenon, although still far from complete, has led to recognition that through atmospheric and oceanic teleconnections, it affects and is affected by numerous, seemingly disparate components of the global climate system.

Mean conditions in the tropical Pacific involve easterly winds that combine with a relatively cool, relatively shallow thermocline to create zonal asymmetry in SSTs, SSH, SLP, and thermocline depth. Along the equator, where Coriolis forcing is zero, water is blown directly west by the easterlies. SSH is about 50 cm higher in the west than in the east. This bulge is known as the western warm pool (WWP). Below it, the thermocline is depressed to a depth of about 150 meters. Temperatures in this pool are the warmest of the Pacific. Convection is generally centered over this region of warm SSTs. Air rises here and diverges both meridionally and zonally into Hadley and Walker circulations, respectively. In addition, convective zones mark ribbons of convergent activity – the more equatorially situated ITCZ and the southerly, off-equatorial SPCZ. Trade winds from both the northern and southern hemispheres converge on the ITCZ. Due to asymmetric heating of hemispheres due to landmass configuration, the ITCZ is always positioned slightly north of the equator. This positioning draws the SE trades across the equator, transporting negative vorticity into the region, intensifying convective activity further. In addition, as the southeasterlies blow toward the equator, Coriolis forcing to the left of travel causes an Ekman divergence of surface waters to the south. Along the equator, surface water is blown directly west, and across the equator, in the northern hemisphere, waters diverge to the right. The consequence of these various forcings within a relatively narrow region is the strong development of a cold tongue – a region of intense upwelling of cool waters from below a highly shoaled thermocline. With the exception of a strip of warm surface waters that extends across the ocean basin just north of the equator, SSTs are lower in the east than they are in the west.

Westward flowing surface currents – the North and South Equatorial Currents (NEC, SEC) transport the cold water westward. The NEC is slow and broad, while the SEC contains a swift jet within it. An eastward flowing surface countercurrent, the North Equatorial Counter Current (NECC), follows the band of warm surface waters mentioned above. Its flow is to the east, this, despite the fact that the prevailing winds are easterly. This current responds to gradient flow due to the sea-surface gradient forced by the prevailing wind flow. The NECC is positioned between 3°N and 10°N. It is a narrow, fast current. There is also a South Equatorial Counter Current (SECC) at about 9°S; it is transient, becoming increasingly so east of the Date Line. Deflection from the flow of the NECC is to the right of travel, toward the equator. This, along with the flow of cross-equatorial southeasterlies, contributes to water convergence at about 4°N. This marks a latitudinal high point in the SSH. As a result of convergence and Ekman pumping, downwelling occurs here. The thermocline below is depressed. Flow at depth is toward the equator. From about 4°N northward, the sea surface slopes downward, the thermocline upward. The height along the sea surface between these latitudes changes according to season and ENSO phase and plays a significant role in magnitude and direction of oceanic heat and mass transport. Flow down gradient at depth along the equator is eastward. This is the Equatorial Undercurrent (EUC). Much of the water that feeds this subsurface current above the thermocline is from the Southern Hemisphere. Upwelling along the equator at the surface diminishes transport of the EUC downstream. This current is transient, known to disappear during strong El Ninos. Summarizing meridional motion in the tropical Pacific: Ekman transport of surface water is poleward and geostrophic transport is equatorward at the depth of the thermocline.

The equatorial Pacific hosts a strong seasonal cycle of winds. Oscillatory behavior of the ENSO system is superimposed upon it. During the boreal summer SE winds are strong; zonal asymmetry along the equator is at a maximum. The ITCZ is displaced to its maximum northward position. As insolation shifts with the seasons, boreal fall into winter, SE winds abate. The ITCZ shifts southward. The Hadley cell weakens, wind stress along the northwest coast of South America weakens. Upwelling of cool waters in the east-equatorial Pacific (EEP) slows and SSTs in the east rise as a consequence. As the zonal asymmetry of the equatorial Pacific weakens, SE winds continue to abate in response. Intense convection moves southward and eastward, strengthening the SPCZ and weakening the ITCZ over the western Pacific, drawing both zones closer to the equator – the ITCZ southward and the SPCZ northwards - the SPCZ following the warm pool. This stage of the seasonal sequence leads to de-stabilization of the coupled system. Conditions are *vulnerable* to an El Nino event in the early boreal winter.

Out-of-view during this abatement of wind strength and relaxation of the SSH gradient is the occasional propagation of planetary-scale gravity ocean waves along the equator. Change in wind strength and/or direction generates equatorially confined Kelvin and slightly off-equatorial Rossby waves that travel along the thermocline, affecting its depth and providing the ocean with "memory", allowing anomalies resulting from atmospheric changes to persist beyond their short-lived forcing.

These waves tend to both amplify, and later eliminate, consequent changes in SSTs, SSH, SLP, and wind strength. These ocean long waves, discussed in greater detail shortly, can, but don't always, lead to an El Nino.

After reaching a minimum during the boreal spring, if an El Nino event was not triggered, the SE trades begin to resume their strength. The ITCZ migrates northward and strengthens with the migrating insolation. Upwelling of cool waters intensifies in the east. When boreal summer returns, the ITCZ further strengthens, while the SPCZ weakens. Both convective zones move away from the equator. Zonal and meridional convection patterns follow the SSTs.

If seasonal strengthening of easterly winds fails and westerly wind anomalies in the central and western Pacific get the upper hand, the system may fall victim to a warm event. How, exactly, the system "falls victim" to an event is not fully understood. At times a trigger is apparent. At other times, no such trigger is evident. But when it does happen, zonal asymmetry is minimized, temperatures in the west cool due to eastward advection of warm water, while temperatures in the east warm, largely due to reduced evaporation as a consequence of reduced winds. The ITCZ fails to move northward. The SPCZ merges with the ITCZ. The Walker Circulation shifts to the east and weakens. The Hadley cell re-configures, contracts, and intensifies, enhancing heat transport poleward, especially into the Northern Hemisphere. The NH subtropical jet intensifies over the Pacific Ocean and over the continental U.S.; it weakens over the Atlantic. The polar jet is strengthened. Storms tracks are re-routed.

During an El Nino, two upper-level anticyclones, straddling the equator near the date line, become established. This anticyclonic couplet, resulting from anomalous heating of the atmosphere, represents upper-level divergence, over the low-level convergence – the area of enhanced convection and precipitation at the surface. These upper-level anticyclones are associated with a pronounced teleconnection pattern to higher latitudes, as manifested in a train of highs and lows (positive and negative geopotential height anomalies) originating at an anomalous heat source in the central tropical Pacific and continues throughout the upper levels of the atmosphere over the northern Pacific Ocean and North American continent, curving eastward and equatorward over the eastern United States. This is the Pacific-North American pattern – the positive PNA pattern. Upper-level easterlies bound the equatorward side of the anticyclones. These overlie the anomalous westerlies at the surface. On the poleward side of the anticyclones are anomalous upper-level westerlies. These amplify the subtropical jets at the upper levels, displacing them equatorward, particularly in the winter hemisphere. It has been suggested that these anomalous upper-level westerlies that intensify the jets result from the El Nino-intensified Hadley Circulation, as its intensification increases the poleward transport of angular momentum in the upper troposphere.

In the ocean below, Kelvin and Rossby waves have been excited. These ocean long waves – hundreds to thousands of kilometers long – do not always lead to an El Nino, but an El Nino does not occur in their absence. As stated earlier, they are generated from a perturbation, usually a change in wind strength or wind direction. The effect of such a change is the release of the invisible "wind wall" that holds water along the equator to the west. When this wall is eliminated and the water moves eastward, this forces a bell-shaped bulge downward in the thermocline below. Waves move zonally from this point, both eastward and westward. Downwelling Kelvins propagate to the east, typically in the presence of westerly winds, depressing the thermocline along its path.

Upwelling Rossby waves, dominated by easterlies from the point of perturbation, travel westward, shoaling the thermocline along the way. (Relative wind direction determines if the wave is downwelling or upwelling.) It takes about 90 days for the Kelvin wave to make it across the Pacific from its point of origin. Warm SSTs in the central equatorial Pacific delineate the path of the progressively deepening thermocline eastward. These warm SSTs result from zonal redistribution of water from the west and from the further deepening of the thermocline, which reduces entrainment of cold, upwelled subsurface water. These warm SSTs are born of a different cause than the early signs of warming in the east; early anomalies resulted from weakened easterlies. These later-occurring warm anomalies result from gravity waves. When the Kelvins “hit shore” on the western coast of the Americas, coastal Kelvins migrate north and south. Some of that energy is redirected into reflected slightly off-equatorial downwelling Rossby waves, waves whose velocity off-equator is less than a third of a Kelvin wave. Rossby-wave velocity decreases with increasing latitude. Latitudinal variations in velocity of off-equatorial Rossby waves, plus slight latitudinal shifts due to requirements imposed by conservation of vorticity, affect the heat and mass budget of the tropical region. Meanwhile, back to the perturbation site, upwelling equatorial Rossby waves traveling along the thermocline to the west reach the coast of the maritime Indonesian continent. They reflect back to the east in the form of an upwelling Kelvin. The interaction of these and other crisscrossing waves results first in the amplification of the Bjerknes-driven SST and wind changes and then allows the amplification to decay, leading back to a neutral state, or, if the system overshoots, a La Nina state[19-20].

A model that works well to describe many of the observed features of ENSO is the delayed-oscillator model. While not perfect, it combines both the coupled ocean-atmospheric feedbacks of the Bjerknes responses and the interactions of the long ocean waves traveling along the thermocline.

During an El Nino event, heat is lost through atmospheric and oceanic processes. Heat gained by the system during non-El Nino stages is zonally redistributed during an El Nino and meridionally lost at various stages of a warm event. In the atmosphere, heat is lost through changes in convective activity – a weakened zonal Walker Circulation and an intensified meridional Hadley circulation. These changes result in increased latent-heat loss due to intensified evaporation, increased albedo due to changes in cloud cover, and increased outgoing-longwave-radiation (OLR) due to meridionally contracted areas of convection and expanded regions of subsidence[8] within the off-equatorial region of the tropics. Heat removal through ocean dynamics includes decreased delivery of heat from subtropical gyre exchange, coastal Kelvin propagation, cessation of tropical instability waves (TIWs), and increased poleward Ekman transport of warm surface water[21-24].While the atmospheric changes contribute significantly to the tropical heat budget, they collectively have less impact on heat export than do changes in ocean dynamics.

At the end of an event, sea-level data indicate that water mass has not only been redistributed zonally, but meridionally, as well. Both heat and mass are exported from the tropics. This meridional transport is not symmetrical about the equator. More is transferred to the Northern Hemisphere than to the Southern Hemisphere during an El Nino.

Invoking the roles played by Kelvin and Rossby waves, *Jin’97 and An and Kang ’00* suggest that total heat content adjusts by modifying zonal and meridional ocean mass flux (discharge-recharge oscillator model) [25-26]. During a non-El Nino event, net heat transport is equatorward, recharging the system. During an El Nino event, poleward heat transport dominates, discharging the system's heat inventory.

Expanding further on this theme, Brady [24] suggests the tropical heat inventory can be broken down into two dominating systems: a shallow meridional overturning cell (MOC) and the subtropical gyre exchange (GE). The shallow MOC involves cool, deep geostrophic flow equatorward from off-equatorial locations, compensated by poleward Ekman transport of warm surface water along the western and central regions of the tropical Pacific. The GE involves diabatic heating of the cold surface waters of the cold tongue, some of which are subducted downwards and westwards, and some of which are advected westwards by the NEC and SEC. These warm subsurface waters are geostrophically delivered to the EUC. Balancing the import of heat from the GE is the upwelling of cool water from below the thermocline in the east and subsequent transport poleward through the Ekman process. In addition to these two major transport vehicles, tropical instability waves (TIWs) also play a role. TIWs – large wavelength atmospheric disturbances - work along the periphery of the cold tongue in the east equatorial Pacific to mix warmer off-equatorial surface waters into the cooler equatorial waters during non-El Nino events, in essence, bringing heat into the system during the cool phase, competing with Ekman transport of heat poleward. Balances shift among these multiple processes as wind strengths change and SSH gradients adjust. These shifts determine whether transport of heat is dominantly equatorward or poleward.

In summary, at the end of a warm event, ocean heat and ocean mass have been lost from the equatorial system. This has been accomplished through the action of equatorial and coastal Kelvins, the disappearance of TIWs, the weakening or cessation of the EUC, and the augmentation of Ekman transport due to intensified easterlies at the end of an event, when SSTs are at their warmest and cover the most zonal area. In addition, heat has been lost through the atmosphere with increases in albedo, outgoing longwave radiation (OLR), and latent-heat loss. Timing of these heat-transport changes is complex and episodic. Complicated differences in interactions of geostrophy and Ekman transport between the Southern and Northern Hemisphere result in the observation of mass and heat transport going mostly northward, into the Northern Hemisphere, at the end of an El Nino event[27-28, 24].

It is clear from the description of heat gain during the non-El Nino stages and episodic net heat export during the various stages of an El Nino event, that heat content must be a pre-condition for an El Nino to occur. Thus, instead of viewing El Nino in isolation, it must be considered as one of two steps required in the ENSO system – step one being La Nina as the preparation or heat-accumulation stage and step two being El Nino as the heat-export stage. *Sun '04* suggests with his heat-pump hypothesis[23], that in order for the heat-accumulation/heat-export steps to proceed, there must be a pre-existing temperature contrast between the warm pool in the western tropical Pacific and the thermocline water in the equatorial undercurrent (EUC). This requires a meridional temperature contrast, as well. Either the equator must gain extra heat or the subtropics and high latitudes must cool. This allows for a temperature contrast between the upwelled water from below the thermocline and the water in the warm pool. The more significant the temperature contrast, the stronger the La Ninas and El Ninos will be [23]. The end result of this heat accumulation and subsequent export is that through ENSO, the time-mean stratification of temperature between the warm pool and the EUC is maintained. The tropical Pacific retains extraordinary stability, despite thermal forcing. And, with the understanding that the regional dynamics of the tropical Pacific wield strong influence on global climate patterns, one is led to infer from this observation that the tropics play a large role in global climate stability.

It is significant to note, that while the interaction of long waves in the ocean provide memory to the ENSO system, allowing an El Nino event to endure for 12 to 18 months, because Rossby-wave velocity is fastest at the equator, and decreases with distance from the equator, memory in the tropical ocean is shorter than memory at higher latitudes. Thus, this fact allows the tropical ocean to respond relatively quickly to large-scale changes in wind patterns, allowing for a comparatively short adjustment time to atmospherically imposed perturbations[18]. (In contrast, "memory" of high altitude *atmospheric* dynamics is long in the tropics and shorter at higher latitudes[72].)

Paleo-history and More Recent History of ENSO:

It is not a trivial matter to note that El Nino has changed character – its frequency and strength – over time. Some have suggested that the Pliocene – about 4 mya – hosted a weak, but permanent El Nino[29]. Temperatures were, on average, about 3°C warmer than today – much of that increased warmth found at high latitudes. Sea level was about three to five meters higher than today. But this does not describe the El Nino of today. That system of the Pliocene, if it has been accurately assessed, was hardly variable. How did the ENSO system become variable, alternating between La Nina and El Nino, as it does today?

That journey involves tectonic orchestration of oceanic dynamics, and in turn, atmospheric dynamics. Several key pieces play a role in the assembly of this puzzle. They include the Indonesian maritime continent, the Bering Strait, and the Panamanian Seaway. Key variables that were affected by the movements of these pieces include basin width, basin "valves", and subsurface water temperature[29-37].

Critical to the evolution of the modern-day El Nino are three basic parameters in the equatorial Pacific region: a certain size basin width, a shallow thermocline with cold deep waters, and the existence of prevailing easterly winds in the tropics. The story of how this set of conditions was attained in the equatorial Pacific involves at least a few million years. In brief, it included: Tectonic shifts and emergence of landmasses within the Indonesian maritime continent and consequent changes in water source feeding the thermocline and changes in atmospheric convection patterns; the opening of the Bering Strait; and possibly the closing of the Panamanian Seaway, which may have been responsible for both a reversal in flow direction through the Bering Strait to dominantly north and for the re-organization and intensification of deep-water formation in the North Atlantic, thereby enhancing the thermohaline-circulation component of the Atlantic sector's MOC. These tectonic rearrangements worked together over a period of a couple of million years, re-designing oceanic and atmospheric circulation. In the end, ice-sheet growth in the high latitudes of the Northern Hemisphere accelerated. High-latitude waters cooled. Obliquity was low, supporting the ice and temperature changes. The deep waters of the global ocean, fed by the high-latitude waters, cooled[26]; the thermocline shoaled. Strong easterlies developed. The Walker Circulation intensified shortly thereafter, about 1.75 million years ago. A cooling planet, with a shallow equatorial thermocline, with an equatorial basin width of a size, such that interference between equatorial planetary-scale gravity waves allow for the growth and decay of ENSO phases, set the stage for, what some would say, became a Pacific tropical control on global climate[29-38].

Since the time of the inception of the variable ENSO system, it has changed character – at times alternating frequently; at other times, alternating infrequently, or possibly not at all. While speculation is that the variable ENSO system has operated for a million or more years, it cannot be confirmed. Proxy records support the assertion that variability within the ENSO system has operated for at least the last 130,000 years[39]. Mean global temperature appears to display little correlation to ENSO variability of the past[39].

While a damped El Nino expression appears to have dominated much of the glacial period, 21 to 17 kya, during extreme cold of the last glacial, El Nino events increased sharply. Toward the end of the last glacial, the newly initiated warming trend failed. Temperatures reversed and plummeted once again during the Younger Dryas, about 11 kya. Strong El Nino events occurred during this time. After the cold temperatures lost their grip during the early stages of the Holocene, when warmth was quite pronounced, variability of the ENSO system was low to absent. Variability remained minimal during the Holocene Optimum, about 7 to 5 kya[41], when temperatures were about 1° to 2° warmer than today. After 5kya ENSO frequency slowly increased [40]. By 3ka, frequency was similar to today[39]. And more recently, during parts of the Little Ice Age, spanning the 14th to 19th centuries, low activity was punctuated with strong El Ninos. No strong El Ninos occurred between 1579 and 1719, yet powerful ones occurred in 1720 and 1728. These events rival the large one in 1997/98. Another long break in extreme activity occurred between 1728 and 1878[42]. Since this time, there have been seven significant warm events and numerous smaller ones.

It becomes apparent from the record that ENSO activity is not highly correlated to global mean temperature over the paleo record. Some suggest that orbital parameters dictate the ENSO response. One camp points to obliquity [43]; another points to precession [41]. If, as *Sun '04* [23] suggests, either increased forcing in the tropics or decreased forcing at higher latitudes would be sufficient to set the stage for an enhanced ENSO system, numerous scenarios could be invoked to satisfy those requirements. Other mechanisms that have been suggested include explosive volcanic activity[44] in the tropics, variability within the summer Asian monsoon activity [45], perhaps related to obliquity patterns, and ITCZ placement due to changes in the global MOC and ice cover [46].

Is there a pattern? In effort to establish a profile of warm events, a composite El Nino was compiled[47] based on El Nino events between 1949 and 1976. This composite is considered the "canonical view". During this time period, activity was low. According to this view, a build-up period of anomalously strong easterlies precedes a warm event by about a year. Anomalous westerlies supplant the easterlies in the fall preceding the onset of a warm event. An event typically begins at the end of the calendar year, with SSTs peaking along the South American coast during the boreal spring of an El Nino year. Warming both emerges separately in the central Pacific (CEP) and spreads westward from the east after the initial eastern warming.

Nature threw a curve ball the year these observations ended. Events subsequent to 1976 have followed a distinctly different evolution from this canonical view. Between 1977 and 1996, coastal warming peaks occurred in the boreal spring subsequent to an El Nino year rather than of the El Nino year. In addition, the warming occurred more along the Central American coast than along the South American coast. The westward spreading pattern was distinctly different, as well. Throughout much of the 1990s, a low-level, somewhat perpetual El Nino dominated the region; yet confusingly, not all indices indicated a warm event. With the 1997/98 warm event – a particularly large one, there was another change. Initial warming occurred in both the central and eastern Pacific along the South American coast at the same time. In the 2002/03 El Nino, warming started earlier in the year, originated in the CEP, and remained in the CEP.

Federov & Philander '00 [48] suggest that one can see a distinct interdecadal variability of ENSO behavior, oscillating between periods of strong variability and periods of weak variability. They attribute such changes to changes in background state, citing depth of thermocline and time-averaged easterly wind stress as the two critical parameters. Increasing thermocline depth and/or decreasing easterly wind stress have stabilizing effects, leading to reduced variability.

Employing a related line of thinking in attempt to determine the reason for changes in modern ENSO behavior, researchers looked to the source of equatorial thermocline waters for a clue about temperature of the subsurface waters. Source water for the thermocline has historically come mostly from the Southern Hemisphere (SH), particularly from the Antarctic Intermediate Water (AAIW) and subducted waters of the subtropical gyres (twice as much from the SH gyre). Incorporated within the AAIW is water from the Antarctic Circumpolar Current (ACC); thus, intimately linking Antarctic conditions with the tropical Pacific. Researchers speculated that after 1976, source water for above the thermocline might have altered. Tracers gave conflicting results. Increased C-14 tracer values suggested that source water for the equatorial thermocline (EUC) had changed to include more Northern Hemisphere water. This could be explained by a warm anomaly subducted in the late 1960s and early 70s in the North Pacific. Yet salinity and temperature measurements pointed to the warmer, more saline thermocline waters of the SH as the continued source, leading to speculation that this source had perhaps increased. But that conclusion does not coincide with the C-14 data, with slightly more C-14 in the NH. Thus, evidence of a shift in source remains inconclusive[49]. More definitive evidence was found regarding ocean transport velocity. Transport velocity was examined in the 1990s by *McPhaden&Zhang '02* [50]. They found that the strength and volume transport of water was significantly reduced in the subtropical gyres of the Pacific during much of this decade. Temperature of the subsurface water increased accordingly. Once again, the system defied constraints of paradigm behavior. Toward the end of the '90s, the velocity slowing trend reversed. It has since rebounded to conditions seen prior to 1976. [51]

Thus, it is clear, each El Nino is distinct, its background conditions inconsistent. Its history of development, its variable nature in both expression and timing, makes its prediction elusive. Once an El Nino has started, prediction of its development is facilitated. And generally, because of the premise of discharge and recharge of heat and mass required to prime the system, one is somewhat safe in not predicting the occurrence of an El Nino in successive years (the '90s gave challenge to this notion though).

It seems Philander was indeed on to something with his earlier-mentioned encapsulation of ENSO: "the ENSO system is best left to ambiguous and qualitative description, much akin to the term "winter", best understood in retrospect and far from a predictable quantity." ENSO teleconnections follow a slightly less vague pattern. Despite erosion of the once-accepted paradigm of phase-locking of warm events to the annual cycle, there does exist a pronounced seasonal cycle of ENSO teleconnections. Teleconnected patterns move zonally and meridionally. Through these communications, ENSO can remotely influence patterns and other patterns can remotely influence

ENSO. In the zonal direction, changes in SST distributions shift zonal convection centers, providing a tropospheric bridge for transmitting climate anomaly signals. Meridionally, the pathways are more complicated. In some cases, planetary-scale waves are propagated from deep convective activity in the tropics. These waves propagate not only meridionally; some propagate with a large vertical component. The former affect climate patterns more directly. The latter affect climate patterns through interactions with the stratosphere and/or with the polar vortex. Much remains unknown, but much progress has been made in recent years to make sense of all the oscillatory patterns and their interrelationships. A discussion of these interlinked patterns follows. An attempt is made to extract from a collection of seemingly random oscillations, patterns that reveal a set of organized feedbacks that work to maintain a flexible range of climatic stability on a variety of timescales.

COMMUNICATION OF SIGNALS:

The section begins with a look at the Tropics of all oceans – the Atlantic, Indian Ocean, and back to the Pacific.

Tropical Atlantic Variability (TAV):

Two modes of variability exist in the tropical Pacific – zonal and meridional. Due to the dominance of the ENSO pattern, this latter variability in the Pacific is overshadowed. The Atlantic is similar to the Pacific. It, too, hosts both forms of variability. In the Atlantic, both modes of variability are evident, with the interhemispheric being the more dominant. *(Slide 7)*

The interhemispheric, or meridional pattern of variability, once incorrectly thought to be a dipole pattern, results from the difference in SSTs between the tropical northern Atlantic (TNA) and the tropical southern Atlantic (TSA). Because TNA SSTs and TSA SSTs vary independently from one another, this relationship cannot be classified as dipole behavior. Large contrasts of TNA minus TSA SSTs constitute a positive gradient. Such a condition prevailed between 1966 and 1970, 1976 and 1983, and the early 1990s. Mostly negative conditions occurred from 1971 to 1975 and 1984 to 1989. The gradual variability of this pattern represents a phased interaction between unrelated decadal variations in separate hemispheres and is positively correlated with the strength of the Atlantic Hadley circulation, the former lagging the latter by one month. Air rises over the warm SSTs, flows southward aloft into the Southern Hemisphere, and returns toward the warm NH temperatures in the lower atmosphere. When the gradient is positive, this flow is amplified. Return flow of the lower limb of the Hadley is deflected to the west in the Southern Hemisphere and to the east in the Northern Hemisphere. Thus, when the gradient is positive, anomalous winds of opposite sign prevail. Anomalous westerlies act to warm TNA SSTs further, explaining the lagged correlation.

Equatorial variability in the Atlantic resembles that in the Pacific. During an Atlantic warm event, within the Atlantic, the Walker circulation weakens, the Hadley circulation increases, SSTs in the east rise, SE trades subside, and the Atlantic ITCZ migrates southward. Atlantic Ninos occur more frequently and are subdued in parameter extremes and duration compared to their Pacific counterparts. This is in large part due to a narrower basin over which the long ocean waves interact.

It has long been suspected that an El Nino in the Pacific affects tropical SSTs in the Atlantic. During the mature phase of an El Nino, the Pacific Walker cell shifts eastward. The Atlantic responds; yet the responses are inconsistent. Sometimes the Atlantic equatorial response is cooling; at other times, it is warming. In either case, the response lags the El Nino peak by about five months [52-53.] This observation implies that an El Nino can modify intrinsic variability within the Atlantic basin; yet the mechanism is fragile, vulnerable to interference. This remote forcing exhibits a lag of about five months. A look at the specific remote influences of ENSO on the Atlantic might lessen the mystery. There are three remote mechanisms that can warm SSTs of the TNA – a weakened Atlantic Hadley due to shifted circulation, teleconnections through the PNA, and teleconnections through a "tropospheric bridge". *(Slide 8)*

Due to the shift of circulation patterns due to an El Nino, an anomalous pattern is established over the tropical Atlantic. Most pronounced is a weakening of the Hadley circulation in the Atlantic. Anomalous ascending motion over the North Atlantic subtropical high leads to a relaxation of the NE trades, a slowed subtropical gyre, reduced evaporation, upwelling, and entrainment, and thus leading to a warming of SSTs several months later in the TNA in late boreal spring and early summer [52].

Alternatively, El Nino can affect the TNA through the alternating high and low geopotential height anomalies of the Positive North American pattern (PNA), weakening the NH subtropical high, which would slow the NE trades and, due to increased evaporation, increase the TNA SSTs. In addition, a reduction in strength that might occur in the subtropical gyre as a result of El Nino teleconnections, could lead to an increase in temperature of the water delivered to the equatorial thermocline. Positive PNA patterns are often, but not always, associated with an El Nino.

During an El Nino, atmospheric Kelvin waves travel eastward through the troposphere from the eastern Pacific to the Atlantic. Moist convection in the Atlantic is suppressed. Latent heat builds up in the boundary layer. Evaporation, upwelling, and entrainment of cool subsurface water decrease. SSTs increase. In this manner, through what has been called a tropospheric bridge, the El Nino exerts additional warming influence on the Atlantic regime.

These remote forcings do not consistently manifest. The key to that inconsistency appears to lie in the western tropical Atlantic winds. If easterly winds are strong in the equatorial Atlantic, an Atlantic Nino fails to evolve. The irony here is that an El Nino will often generate anomalous easterlies in the WEA. Thus, a Pacific El Nino teleconnects opposing mechanisms to the Atlantic[53]: warming mechanisms, such as the tropospheric bridge, and a cooling mechanism, strong easterly winds, which destructively interfere with the tropospheric bridge. If the easterlies can remain strong, they can thwart an Atlantic Nino. It is unknown what might allow the easterlies to out compete the warming mechanisms. It is equally unknown what causes the easterlies to decay.

Some suggest warm TNA SSTs preceding an El Nino could set the stage for favoring the tropospheric-bridge warming, causing the easterlies to decay. Could the NAO play a role?

The North Atlantic Oscillation (NAO), a topic discussed later in this paper under the umbrella of annular modes, correlates negatively with temperatures in the TNA[52]. There is no similar correlation with temperatures in the TSA; thus, no significant correlation exists with the meridional gradient variability. During a positive NAO, easterlies in the eastern tropical Atlantic are strong and SSTs cool. Opposite conditions prevail during the negative NAO phase. *(Slide 10)* The NAO was positive, with cool TNA temperatures and strong easterlies throughout most of the period between 1972 and the mid-late 1990s. One of the strongest El Ninos occurred during this time – the '82/83

El Nino. An Atlantic Nino did not develop. The easterlies "won" over the teleconnected warming mechanisms. A similarly strong El Nino occurred in 1997/98[53]. The NAO phase had switched to negative. Weakened easterlies and warmer TNA SSTs prevailed. Easterlies gave way to teleconnected warming. An Atlantic Nino developed. It might be further noted that the positive PNA pattern that sometimes teleconnects warming of El Nino to the Atlantic did not develop during the ''82/83 El Nino. A positive PNA pattern did accompany the '97/98 El Nino.

Moving to the other side of the Pacific…

Southeast Asian Indian Monsoon (SAIM):
In the 1920s, Gilbert Walker focused his attention on the occasional failure of the Southeast Asian Indian Monsoons (SAIM). He found that failures typically occurred at times when the sea-level pressure differences between the Indian Ocean and the Pacific Ocean were low - a situation we now know correlates with a warm ENSO event. An eastward shift in the Walker Circulation during a warm phase positions the subsiding arm of the circulation pattern over Asia, suppressing the monsoon – thus explaining the monsoon failures.

This relationship of failed monsoons with El Nino events has recently shown inconsistency. Since 1981, this relationship has weakened. *Kumar '99* [54] has offered two explanations: one, a southeasterly shift in tropical Pacific convection and two, warming temperatures of Eurasia in the boreal winter and spring (related to the positive phase of the NAO). In the former, a shift in tropical Pacific convection could displace suppression of rainfall away from the Indian subcontinent. In the latter, intense warming of the Eurasian landmass due to NAO-related forcings, would allow monsoonal winds to remain strong, even when El Nino dominates, overriding the suppressive effects of a shifted Walker cell. *Kumar et al. '06* [55] concluded that position of the El Nino SST anomaly is critical to monsoonal activity in Southeast Asia. When the anomaly is in the eastern part of the basin, monsoonal activity is unaffected. When the anomaly is shifted more westerly into the central Pacific, as is typical of some of the more recent warm events, monsoon failure is more likely. Reasons for the shifted position of anomaly remain unclear.

Indian Ocean Zonal Mode (IOZM) and the Indian Ocean Dipole (IOD):
Two modes of variability occur in the Indian Ocean; both can be as a result of ENSO, but not always. According to *Yu & Lau '04*, the two modes include the Indian Ocean Zonal Mode (IOZM) (aka: the Indian Ocean Dipole (IOD)) and a basin-wide warming [56].

The IOD, discovered in 1999 by Professor Yamagata, Dr. Saji, and associate of the Climate Variations Program of Frontier Research System for Global Change, is an interannual climate mode occurring in the tropical Indian Ocean. There is a positive phase and a negative phase. During the positive IOD, SSTs drop in the southeastern part of the ocean basin – off the northern coast of Australia, throughout Indonesia, and off the eastern coast of Japan. Concurrently, SSTs increase in the western part of the basin – north of Madagascar to the northern edge of Somalia. Convective activity increases over North Africa, over India, and off the eastern coast of Africa. During the opposite phase – the negative IOD – conditions are reversed: cool in the west, warm in the east. Convection centers hover over Australia, Indonesia, and Japan. *(Slide 13)*

Model studies indicate that, while ENSO can force the variability, variability can exist without the presence of ENSO. This suggests that ENSO is capable of changing the dominant timescale of the

IOD [56]. Multi-century coral records[57] reveal significant correlation with ENSO periodicities. While debate continues over the role of ENSO, model studies indicate that in the presence of ENSO, the IOZM (IOD) peaks about three to five months after ENSO has peaked. Thereafter, warm SST anomalies in the western Indian basin decay and spread eastward across the entire basin. This basin-wide warming, as generated in the model, peaks about eight months after the IOZM peaks. Then this warm anomaly continues east until a reversed polarity of the IOD develops. With warm SSTs in the east and cool ones in the west, the IOD is in its negative phase. This decays and leads to basin-wide cooling. Speculation on mechanisms for this signal communication suggests that both tropospheric transmission and Kelvin-wave activity are responsible for transporting thermal signatures to remote areas within the basin. (*Slides 13 & 14)*

Teleconnections can operate in reverse, as well. It is posited by *Annamalai et al* (in press) that basin-wide warmings and coolings generate positive and negative precipitation anomalies[58]. The associated convective patterns influence wind anomalies in the western equatorial Pacific. In the case of basin-wide warmings, anomalous easterlies will result in the WEP. If one occurs during an El Nino, development of an El Nino is moderated. The opposite is true for a basin-wide cooling in the Indian Ocean.

Pacific Decadal Oscillation (PDO) = (NPO) ~ (IPO):
Natural variability in background conditions of the equatorial Pacific may govern ENSO behavior. The largest amplitude of interdecadal variability that occurs in the North Pacific is the Pacific Decadal Oscillation. In *1997, Nathan Mantua* [59] identified a basin-wide pattern of SST distribution. In *2001* he described it as "an El Nino-like oscillatory pattern of climate variability centered over the Pacific Ocean and North America". He termed this pattern the PDO [60]. This is the same pattern that Walker had noted in the 1930s and identified as the North Pacific Oscillation (NPO). Mantua found that PDO variations were most energetic in the boreal winter and spring. Differences between El Ninos and the PDO include 1) duration – an El Nino typically persists for 12 to 18 months; PDO operates on two timescales, one 15 to 25 years, the other 50 to 70 years[61], 2) signature – El Nino-associated anomalies are most pronounced in the tropics; the PDO signature is more evident in the Pacific/North American Sector, and 3) cause – causes for El Nino are fairly well known; causes for PDO remain a mystery. *(Slide 16)*

A quasi-symmetric Pacific-wide manifestation of oscillating SST anomalies appears to be a larger feature of which the PDO is a part. This dual-basin behavior, described by *Folland et al. in 2002*, after Mantua's PDO, is referred to as the Interdecadal Pacific Oscillation (IPO) [62]. A PDO warm phase, also known as a positive IPO phase, features a strong Aleutian Low. Its strong counterclockwise winds usher warm SSTs from the tropics, up along the western coast of North America. In the central and northern portions of the basin, SSTs are anomalously cool. A PDO cool phase, or negative IPO, shows a weakened Aleutian Low, a cooler tropical and western coastal regions, and warmer-than-normal central and northwestern basin SSTs. Anomalies extend through the depth of the troposphere and are well expressed as persistence in the Pacific North American (PNA) pattern – a deep Aleutian Low over the Pacific basin and a strong high pressure over western North America[63] El Nino events are often expressed through an enhanced PNA pattern, an observation that emphasizes the often-similar characteristics of the two phenomena. *(Slide 17)*

Empirical studies indicate that ENSO influences on North American climate are strongly dependent upon the phase of PDO. There is reason to believe that frequency and intensity of ENSO phases correlate to PDO phase. An above average number of La Ninas tend to occur during a cool phase of the PDO and an above average number of El Ninos during a warm phase of PDO. An El Nino superimposed on a warm PDO phase will amplify the effects of the El Nino expression. Thus, the phase of PDO appears to modulate the strength and frequency of events. [64a] *(Slide 18)*.

Switches between phases are startlingly abrupt, occurring within less than a year. PDO reversals are accompanied by changes in the behavior of ENSO. A cool regime persisted from 1890 to 1924; it abruptly switched to a warm regime from 1925 to 1946, then back to a cool regime until 1976. A warm regime followed until about '99, when a short reversal to a cool phase occurred. The cool phase appears to have given way to a warm phase again. Possibly it has reversed again; it is not yet known. Short reversals within longer time frames are not uncommon. They occurred within the longer time sets given above, but were not mentioned.

Madden-Julian Oscillation (MJO):
The MJO is a short-term oscillation that appears to play a big role in global climate dynamics. Discovered in 1972 by researchers Roland Madden and Paul Julian, both of NCAR, this multi-spatial-scale feature eludes explanation. In essence, it is an eastward traveling organized collection of convective systems. It circumnavigates the globe every 30 to 90 days, with a maximum amplitude revealed in the Indian and West Pacific Oceans. The cloudiness that illuminates its presence in the Indian and Pacific Oceans dissipates by the time it reaches the South American coast, but the disturbance continues to circle the globe, sometimes erupting into regions of cloudiness. Madden and Julian suggested that this area of convective activity moving eastward in the tropics would disturb the boundary between the troposphere and the stratosphere. It does strongly affect the wintertime jet stream and numerous atmospheric features[65-66]. (*Slides 19-20)*

An MJO envelope, spanning about 10,000 km, contains within it collections or families of superclusters. Each "family" covers about a thousand kilometers. Within the supercluster-family organization are cloud-cluster families – groups of cloud clusters, which are mesoscale convective systems (MCS), each group covering about 100 km. The huge MJO envelope travels eastward, as do the families of superclusters. The cloud-cluster groups within the superclusters travel in opposition, to the west. This net-eastward traveling convectively coupled system affects climate on a global scale through a northward-propagating Rossby-wave train – a series of alternating strong high and low-pressure systems. *(Slide 21)*

The MJO is associated with monsoonal activity in both India and Australia and may play a role in triggering ENSO activity. It is vigorous prior to the peak of an El Nino event. The activity moves eastward with the ENSO SST anomaly[67]. MJO activity is anomalously low after the El Nino event and during non-El Nino events. [68] *(Slide 22)*

Some suggest that without stochastic forcing, an El Nino would not be triggered. Along this line of thinking, MJO stands out as a strong source of stochastic forcing. Does it play a larger role than other stochastic forces – tropical cyclones, westerly wind bursts (those that are not MJO-spawned), wave disturbances? This question remains unanswered.

MJO may affect ENSO by reducing the zonal SST asymmetry. A few mechanisms have been proposed: 1) The MJO can cool the WEP. 2) An easterly zonal current, forced eastward by westerly winds of the MJO, can advect the eastern edge of the western pool. 3) Oceanic Kelvin waves, forced by the MJO, can propagate to the east, depressing the thermocline along its path. All would work to reduce zonal temperature contrasts, setting the stage for a warm event. MJO may well coincide with primed conditions in the tropical Pacific to help initiate an El Nino. Obstructing the clarity of the connection is the difference in periodicity of the two phenomena[67].

MJO also appears to affect annular modes, a topic discussed shortly. In short, high MJO activity is correlated with a high index of the NAM (AO) and SAM (AAO). It is possible the annular modes affect tropical convective activity. Tropical activity, in turn, can affect the annular modes. All this is discussed in the annular mode section.

Quasi-biennial Oscillation (QBO):
Generated by dynamical activity in the tropical troposphere (near the tropopause), the QBO is a mostly stratospheric phenomenon, an oscillation between downward propagating alternating easterly and westerly wind flow in the stratosphere, between 30 and 23 km, over the equator, with a quasi-biennial period of about 27 months. There is a mesospheric QBO and a tropospheric QBO. These are not discussed here. *(Slides 23-25)*

The QBO signal is very difficult to separate from the ENSO signal. In fact, ENSO contains a quasi-biennial component of 22 to 28 months[64b,69]. Both modes tend to reflect phase-locking to the annual cycle; the phase change of the QBO tends to occur during the late boreal spring. Suggested by this observation is the observation that the east and west QBO phases are not equal in duration.

It is believed that upward propagating, internal, equatorial planetary-scale waves drive this oscillation. Rossby waves in the mid-latitudes are known to be deflected vertically by topography and air-mass contrasts. At the equator, due to absence of the Coriolis Effect, the infinite vertical component provides a situation for equatorially generated waves to propagate vertically. These disturbances propagate primarily as internal gravity Rossby waves. Eastward-propagating Kelvin waves can be considered a particular kind of internal gravity wave. Thus, equatorial Kelvin and Rossby waves, along with a variety of other gravity waves, provide the momentum that drives the oscillatory flow of the QBO.

Convective activity is the source of these planetary-scale waves. The MJO is a significant contributor. Easterly or westerly momentum is deposited by these upward traveling gravity, Rossby, and Kelvin waves as they descend, which they do after experiencing decay due to radiative cooling via their upward propagation. The wind regime propagates downward with time. The easterly phase is stronger and persists longer at high altitude than the westerly phase. The westerly phase is more intense, descends more quickly and persists longer at lower altitudes than the easterly phase. Upon descent of these flow regimes, velocity of the flow decreases with decreased height.

There exist critical levels of absorption and shielding effects by descending wave patterns, blocking the penetration by a non-synchronized wave structure. A descending westerly wind will develop a zone of westerly shear. If both atmospheric Rossby and Kelvin waves propagate upward, the westerly Kelvin wave will be prevented by this shear zone from penetrating. Only the easterly Rossby wave can gain entry, making its way to higher altitudes where it, too, will radiatively decay and descend, having imparted its westward-directed momentum flux. The opposite situation is true for the Kelvin to gain entry. *(Slides 25 & 26)*

Through these alternating stratospheric winds, disturbances from the tropical troposphere have the ability to affect the stratosphere, where their influence becomes magnified. Through interactions with the polar vortex, which descends to a tropospheric phenomenon during the winter months, these tropical disturbances are able to affect climate at mid and high latitudes. The mechanism behind this remote transmission is explained in the section to follow, under annular modes. And, while tropical convection, as orchestrated by QBO, plays a role in vortex strength, the QBO, in turn, affects tropical convection. During the QBO west phase, tropical convection appears to be diminished. During the QBO east phase, tropical convection is enhanced, in particular, deep convection.

Mechanisms for these observations may lie in both tropical tropopause height and shear distribution across the tropics aloft. In the QBO west phase, as the westerly winds descend, the wind shear anomalies on the tropical tropopause lead to subsidence, and thus to gradual warming and to a lower tropopause. Distribution of shear is such that it is magnified close to the equator and lessened off-equator. The end result of both these features of descending westerly winds is that convection is shallower and less intense. The opposite is true for the east phase. Deep convection characterizes the east phase of the QBO. Positively feeding back on this behavior, deep convection acts to further elevate the level of the tropopause. These QBO-tropical convection relationships appear to hold in all seasons except summer. The relationship is especially evident in autumn and winter[70-72.] This may partly explain why there exists a QBO signature in the ENSO periodicity, as deep convection is often linked to triggers for a warm episode.

<u>ANNULAR MODES AND RELATED TELECONNECTIONS:</u>
Annular modes behave both independently and intertwined with tropical processes. They, along with ENSO, appear to govern global climate variability, but not without assistance from numerous ancillary oceanic and atmospheric patterns. An examination of these high-latitude dynamics follows. *(Slide 27)*

Low pressure dominates the polar center. An annulus of high pressure encircles it. Pressure systems between the polar region poleward of about 60° and a ring around the 45° latitude oscillate in strength on the order of days. Preferred patterns of strength exhibit decadal variability. This annular mode of variability exists, with slight differences, in both hemispheres. A belt of westerly winds encircles the region at about 55°. The strength changes with changes in pressure gradient between the two pressure regions[73-75].

The oscillation in atmospheric pressure extends through the depth of the troposphere. Westerly winds in the stratosphere circle at about 65°. These winds encircle the polar vortex – a mostly stratospheric phenomenon. During the winter, conditions change; the vortex extends into the troposphere and the westerly winds of both the troposphere and stratosphere couple.

The surface expression of this coupled vortex oscillatory behavior is known as the Arctic Oscillation (AO) in the Northern Hemisphere and the Antarctic Oscillation (AAO) in the Southern Hemisphere. They are also known by their more formal names – the Northern Annular Mode (NAM) and the

Southern Annular Mode (SAM). These terms imply coupling between the troposphere and stratosphere[76-77]. (The NAO is thought to be intimately linked to, if not subsumed by, the AO.)

When the pressure difference between the poles and the mid-latitudes is large, the annular mode is in the positive phase; the vortex is strong and contracted. The jet stream is displaced poleward; storms track at a higher latitude. Temperatures in mid-latitude winters are generally warm. This is due to two factors: one, the westerlies are stronger, and thus draw warm air from the lower latitudes poleward and eastward, and two, frigid air is locked within the strong polar vortex, preventing a chilling outbreak from the polar regions into the lower latitudes. When the pressure difference between the poles and mid-latitudes is small, the annular mode is in the negative phase. The polar vortex is weak and expanded. The jet stream shifts equatorward. Polar temperatures are warmer; yet mid-latitude winters are generally cooler.

Numerous influences can affect vortex strength. Anything that leads to cooler temperatures within the vortex during the winter leads to a stronger vortex. Ozone inventory, greenhouse gases, solar forcing, and volcanic eruptions directly determine the vortex strength through the pressure gradient[78-80]. *(Slide 28)*

As the stratospheric vortex begins to couple with the tropospheric winds in the early day of winter, the vortex is vulnerable to perturbations. Perturbations delivered at this fragile time can lead to a weakened vortex. Perturbations can come from Rossby waves from tropical convection (MJO or ENSO, for examples). Phase of QBO plays a role in how these perturbations interact with the vortex. These perturbations cause stratospheric sudden warmings (SSW) – increases in temperature of 50°C or more within the vortex. This leads to the subsequent erosion of temperature/pressure gradient. If these SSWs occur early enough in the season, the vortex may be able to reform. If not, it remains weak throughout the season. If the vortex has time to re-develop integrity and is hit by these perturbations, instead of the perturbations penetrating the vortex and destroying it, the perturbations can impart westerly angular momentum, actually strengthening the vortex. It is all a matter of timing in how the vortex will react, and thus how the mid-latitude winter climate will evolve. *(Slide 29)*

Direct forcing effects play a large role in vortex strength in both hemispheres. Perturbation effect on vortex strength is an influence largely relegated to the more vulnerable Northern Hemisphere vortex; this is related to landmass distribution and topography, resulting in both thermal and dynamical effects that compromise the vortex's integrity.

The seasonal cycle of the vortex involves stratospheric temperatures. During the hemisphere's winter, the high-latitude stratosphere cools, leading to an intensified vortex, with strong westerly winds aloft. When spring and summer temperatures reverse that warming trend, not only does the stratosphere decouple from the troposphere, but the westerlies in the stratosphere are replaced by easterlies.

In the QBO west phase, the polar vortex is stronger; in the QBO east phase, the polar vortex is weaker. These modulations thereby affect the NAM index. This is done through modulation of the wave guide along which reflected planetary waves travel. In the Northern Hemisphere, Rossby waves reflected off topography or sharp contrasts between air masses propagate vertically and then equatorward. Vertical propagation involves waves with the largest spatial scales – usually wavenumbers one and two. When the waves hit the vortex, they can distort its symmetry or, if the hit is hard enough, can displace it from the pole. If the amplitudes are large enough (or the vortex is not yet strong enough), they might break through the polar night jet that encircles the polar vortex. Upon descent, the waves can cause warming of 50°C or more within the polar vortex. Such events are the SSWs, mentioned above. This causes the westerlies to weaken or reverse, leading the vortex to weaken – a state it may or may not recover from. The planetary waves in the Northern Hemisphere are of a size just large enough to sometimes cause these warmings, and other times not. Thus, there is a large interannual variability in strength of the polar vortex as a consequence. It is this sensitivity to the effects of the vertically propagating waves that allows the equatorial QBO to have an effect on the integrity of the polar vortex, and by association, on the NAM.

If the mean QBO flow in the tropical stratosphere is westerly, the reflected planetary waves can travel into the tropics without encountering any obstacle. But, when there are easterly winds, the equatorward propagating Rossby waves cannot so easily make it to the tropics. Quasi-stationary waves (QSW) are unable to propagate in the QBO easterly winds. Instead, they must travel along the wave guide, which refers to the boundary between westerlies and easterlies in the stratosphere and is the "critical line for waves with zero phase speed". When there are easterlies in the tropical stratosphere, this wave guide is narrow. The end result is that when the QBO is in its easterly phase, planetary waves cannot so easily make their way to the tropics and instead are re-directed to interfere with the higher latitudes where the vortex is. Thus, the wave activity is stronger at high latitudes, leading to greater wave-induced drag on the mean flow of the upper westerlies. And, if the westerlies encircling the vortex are weakened, the region within the vortex warms and the vortex weakens. Thus, the polar vortex is weaker in the QBO-east phase, stronger in its west phase.

Relating this conclusion back to the QBO effect on tropical convection, the relationship becomes clear. When vertically propagating waves are allowed to head to the tropics, as they are during the west phase of the QBO, they thermodynamically lower the troposphere, thus weakening convection there. The higher latitudes escape assault; thus, the vortex remains strong. When the vertically propagating waves do not have easy access to the tropics, as is the case during the east phase of the QBO, they focus their energy on the vortex. The tropical tropopause remains high, inviting deep convection, while the NAM has weakened. In summary, when the QBO is westerly, the polar vortex is strong and tropical convection is weaker. When the QBO is easterly, the vortex weakens and tropical convection is stronger *(Slides 29 & 30)*.

It has already been established that the ENSO and QBO signals are difficult to disentangle. Observations reveal that the AO/NAM weakens during an El Nino by about one to two percent during the maximum SST anomaly [81]. This is attributed to the spawning of planetary waves from deep convective activity due to the SST anomaly. Furthermore, the Arctic center of activity is more prominent and extends more deeply into Siberia during a warm ENSO event. There is little to no center of activity in the Pacific during an El Nino. There is an accompanying equatorward shift of the polar jet stream and associated storm track [82]. Three to four seasons after the maximum SST of an El Nino event, empirical evidence suggests that the vortex strengthens and contracts by two to three percent [52]. During the cold ENSO phase, Quadrelli and Wallace found that the North Pacific center of activity was more pronounced and positively correlated to the North Atlantic center [82].

While ENSO is a tropical phenomenon, it influences wintertime planetary waves at higher latitudes, likely related to the augmentation of the Hadley circulation that occurs during an El Nino. In addition, a strong vortex tends to be mutually exclusive with a strong subtropical jet. During an El Nino, the NH subtropical jet, a thermally driven jet, is strengthened. (An enhanced vortex seems to require a weakened subtropical, thermal jet in order to rev up its eddy-driven polar-front jet – a trade-off of energy exchange.) [81, 74]

No longer are the annular modes being viewed as a strictly high/mid-latitude phenomenon. The signature extends to the tropics. During the positive NAM index, bands of zonal wind anomalies extend to the equator and into the subtropics of the Southern Hemisphere. While it has been shown that tropical convection, and possibly ENSO, influences the strength of the NAM, in turn, the NAM influences the phase of ENSO. Observational evidence reveals a statistically significant relationship between the NAM signature and ENSO[74]. Westerly wind bursts are known to trigger some El Ninos. The positive phase of the NAM is associated with WWBs in the tropical west Pacific. During a positive NAM, anomalous westerly winds develop in the lower troposphere at the equator with an approximate two-week lag in the boreal spring. The winds are flanked on either side of the equator by vortices. Westerly wind bursts channel through these twin vortices. The WWBs are centered over the SST maximum. The westerly anomaly persists through the boreal summer. It broadens to the eastern Pacific by the following autumn. The westerlies cover the entire western Pacific by early winter. El Nino-related warm anomalies become apparent in boreal winter. Model studies were carried out to simulate this observed pattern. Results were consistent with observations [83-84].

Nakamura et al.[83] suggest a mechanism can be found related to the strong polar jet associated with a positive NAM. The circulation pattern accompanying a strengthened vortex involves anomalous ascent at high latitudes, where pressure is unusually low, and strong subsidence at midlatitudes. The subsidence would strengthen the poleward arm of the Hadley cell, amplifying northerly flow in the lower troposphere. This could usher in cold, dry air from off the Asian continent to the western Pacific. The cold, dry air would, in turn, receive sensible and latent heat from the warm SSTs. These processes may generate a twin-vortex structure and associated westerlies, as observed.

A similar two-week lagged coupling between a positive SAM and westerly wind anomalies in the tropical western Pacific are evident; although the correlation is not as strong as it is with the NAM. [84] That could be a matter of timing. As the following discussion reveals, these relationships and teleconnections are not always consistent.

ENSO's relationship with the SAM affects ENSO's teleconnections to the Southern Hemisphere. Moisture flux in West Antarctica appears to be correlated with the phase of ENSO. El Nino, through the PSA pattern (analogous to the PNA), increases moisture flux in the Western Antarctic region. This pattern was particularly evident between 1990 and 2000. But this relationship is not always apparent. There appears to be a decadal variability in the robustness of this teleconnection. When the SAM and ENSO phases are strongly correlated and coupled, ENSO teleconnections are strongly communicated to the high southern latitudes. During the 1980s, this teleconnection was weak. After 1990, the teleconnection strengthened. After 1990, significant shifts in tropical Pacific convection occurred. The PSA pattern was observed to strengthen[85]. Perhaps this shares a similar mechanism with the PNA/PDO pattern.

The PSA pattern played a prominent role in a precedent-setting event in 2002. In late September of that year, a sudden-stratospheric warming (SSW) lead to the sudden breakdown of the Southern Hemisphere polar vortex. No such event has been recorded in the SAM since observations began in the 1950s. It appears to have been tied in with the PSA. As mentioned earlier, the Southern Hemisphere vortex is stronger and more symmetric than the Northern Hemisphere vortex. This is due to hemispheric differences in landmass configuration and topography; the minimal extent of it in the SH reduces dynamical interactions of planetary wave activity. Without this dynamical wave activity, the vortex escapes impacts that could lead to its weakening. This was not the case in 2002. Late in that year, in mid-September, anomalous deep convection in the SPCZ in the tropical South Pacific spawned a Rossby wavetrain, quasi-stationary Rossby waves (QSW), from which a blocking flow configuration developed. The wavetrain traveled eastward over 21,000 km over a nine-day period. Upward-propagating waves emanated from the ridge that had propagated into the South Atlantic. This wave activity led to the unprecedented breakdown of the vortex. While this was the strongest event of the year, leading to the breaking apart of the vortex, three minor events of similar source and nature preceded this final event, thereby weakening the PNJ substantially. In addition to the amplified PSA pattern that led to this event, unusually strong easterlies occurred in the equatorial upper stratosphere. Recall that when the QBO is in its easterly phase, not only is tropical convection enhanced, but waves vertically propagated from the enhanced convection cannot travel easily back to the tropics; instead, they are often re-directed poleward, where they interact with the vortex. Thus, tropical convection and the amplifying effects of the east phase of the QBO led to a breakdown of the SAM, the collapse of the Antarctic ozone hole, anomalously warmer temperatures over the pole (most of Antarctica), and anomalously cooler temperatures over the western peninsula. Westerlies were less intense both in the stratosphere and the troposphere. When this occurs in the Northern Hemisphere, temperatures in the mid-latitudes are cooler. Tropospheric anomalies persisted for approximately 90 days[86].

The SAM varies on intraseasonal time scales, as well. The Antarctic circumpolar current (ACC) fluctuates on this temporal scale too. *Matthews & Meredith '04* [87] found that during the austral winter, approximately seven days following anomalous MJO convection in the Indian Ocean, anomalous westerlies develop around almost the entire circle at 60°S. This pattern projects strongly onto the SAM, enhancing the westerlies, and thus the acceleration of the eastward ocean current, the ACC. The strength of the SAM westerlies peaks about seven days following the peak of an MJO event and the ACC enhancement follows three days after that *(Slides 31- 34)* .

Antarctic Circumpolar Wave (ACW):

Van Loon and Shea '85 [88] noted that warm SST and low SLP anomalies were propagating eastward in the western and central subtropical South Pacific Ocean during the year prior to an El Nino event for the years '51, '53, '57, '72, and '82. These anomalies propagated into the eastern tropical Pacific during the following El Nino year. Others[89-90] have observed interannual sea-ice extent (SIE) anomalies in the eastern Indian and western Pacific sectors of the Southern Ocean one to two years before an El Nino onset. Communication between the tropics and the Southern Ocean appears strong. Researchers[91] think they can explain it through the interaction of two globally circumnavigating waves that work together to enhance or damp ENSO phases – the oceanic ACW and the atmospheric GEW.

Identified by *White and Peterson '96* [92], the ACW consists of SLP and SST anomalies that propagate eastward around the Southern Ocean in about eight to nine years with a zonal wavenumber of two; thus, its apparent period at any location is about four years. It spans the extratropical Southern Ocean from around the sea-ice edge at about 65°S to the subtropical region in the various oceans at about 30°S. Significantly, the ACW does not follow the main core of the ACC, nor does it propagate at the same speed *(Slide 35)*.

The ACW fluctuates on decadal, interdecadal, and multidecadal time scales. This appears to be tied into its thermally derived nature. It exists mostly due to thermal gradients between the subtropics and the sea-ice edge. Changes in these gradients translate to changes in the ACW path.

The ACW cannot be reasonably examined without a co-examination of the GEW – the global ENSO wave. El Nino, with its associated evolution of anomalies of SST and SLP on a global basis, can be considered in terms of a global ENSO signal. This signal comprises a standing mode of covarying SST and SLP anomalies and a traveling wave of covarying SLP and SST anomalies. The standing mode, driven by El Nino, is teleconnected through accompanying changes in the Walker and Hadley circulations, as previously discussed in this paper. The traveling wave of ENSO is a coupled SST/SLP wave that propagates slowly to the east with wavenumbers one and two, taking four to eight years to circle the globe. This is the GEW.

The ACW travels more slowly than the higher latitude GEW, except in the vicinity of 90°E and 150°W. This is the latitudinal region of the warm pool. The GEW slows down during its transit through this section. It slows to a pace that is in unison with the ACW. In addition, the GEW excites quasi-stationary Rossby waves as it crosses the warm pool. These waves phase lock the GEW with the ACW. These meridional atmospheric teleconnections communicate the ENSO signal along the path of a great circle, thus their effect on the Southern Ocean is confined to an area from 150°W to 30°W. In this way, the GEW, through atmospheric communication, reinforces the ACW in the eastern Pacific and western Atlantic sectors of the Southern Ocean. Then, the ACW continues along its path, independent of the GEW, yet carrying its imposed signal throughout the remainder of the Southern Ocean. It finally propagates back to the warm pool by a northward branch that spawns off the main wave. The ACW and GEW continually and simultaneously reinforce each other, allowing anomalies in both the tropics and in the mid-to- high latitudes of the Southern Ocean to communicate with one another[93] *(Slide 36)* .

It has been recognized that both the ACW and El Nino evolutionary patterns underwent distinct changes around 1977 – the year of the PDO/IPO shift). This has been explained by White and Annis[91] by changing temperatures in the Indian and Pacific Oceans. Prior to "the shift", the subtropics were warm; the tropics and Indian Ocean cooler. After "the shift", SSTs in the central tropical Pacific Ocean and in the Indian Ocean increased. Cooler SSTs in the western and central subtropics accompanied the change. The ACW, dependent on thermal gradients, expanded equatorward prior to 1977 during the warmer subtropical period and retracted poleward after, during the cooler subtropical period. This changed how anomalies were transmitted to the equatorial Pacific.

Prior to 1977, warm anomalies propagated northward along the west coast of South America to the equator [93-94, 88]. After 1977, the tendency of development was for a slow eastward propagation of the anomaly to move from the warm pool. It has been determined that prior to 1977, the path taken by the ACW and its northward-spawned branch, differed from that taken after 1977. The wave was pronounced in the Pacific sector prior to 1977, following the line of warm SSTs in the subtropics, avoiding the cooler Indian Ocean. After 1977, the ACW followed along the Indian Ocean sector of the Southern Ocean and spawned a branch that traversed northward to the equatorial region of the Pacifc, whose subtropical temperatures were now cooler.

Changes occurred in the GEW too. Prior to 1977, the warm phase of the GEW propagated only as far as the central equatorial Pacific. After 1977, the warm phase of the GEW extended all the way to the east equatorial Pacific. It appears that prior to 1977, the GEW signal contributed to the strengthening of the El Nino events. After 1977, the GEW appears to have coincided with, and likely contributed to the initiation of, El Nino events.

These observations tie together anomalies in temperature in the Indian Ocean, Pacific Ocean, and Southern Ocean, anomalies in the latter being transmitted to the tropics according to paths dictated by anomalies in the Indian and Pacific Oceans, and show that tropical dynamics transmit anomaly signals and reinforce the high-latitude wave pattern that enables the combined anomaly pattern to be carried through the Southern Ocean and back to the tropics, where it can again affect the high latitudes!

An added twist linking high latitudes of the Southern Ocean and the tropical Pacific involves the Antarctic sea-ice dipole[90]. Location of regions of expanded SIE around Antarctica appears to be correlated to phase of ENSO. The regions of expanded ice cover are opposite in opposite phases of ENSO. During El Ninos, ice extent increases in the Atlantic sector. This includes the downwelling region of the Weddell Sea. Ice extent decreases in the Pacific and Indian Ocean sectors during an El Nino – the Ross Sea region. The opposite is true for La Ninas. This redistribution of Antarctic sea ice is explained through dynamics of the subtropical jet during different phases of ENSO. During an El Nino, the subtropical jet is intensified in the east Pacific Ocean sector and displaced equatorward. It extends east over the continental U.S. Over the Atlantic, the jet weakens, due to anomalous subsidence resulting from the eastward shift of the Walker Circulation. Where the jet is strong, it diverts storms equatorward, away from the storm intensifying source – cold air from Antarctica. Thus, during an El Nino, when the Pacific sector's subtropical jet intensifies, there are fewer intense storms in the Pacific and Indian sectors in the mid to high latitudes of the Southern Ocean. With fewer storms, there is less wind, and correspondingly, less sea ice. In the Atlantic, with the weakened jet, storms are not diverted. Thus, increased winds within the Atlantic sector of the Southern Ocean that occur during an El Nino result in sea-ice extent broadening in the Weddell Sea region[95]. *(Slides 37 & 38)*

Brine rejection, consequent of sea-ice formation, significantly contributes to the increased density necessary to promote deep-water formation. Both the Weddell Sea in the Atlantic sector and the Ross Sea in the Pacific sector are regions of deep water formation contributing to the Antarctic Bottom Water (AABW), the former being slightly more dominant. How the ENSO phases and their dipole effects on SIE might affect deep-water formation is a topic worthy of investigation.

THERMOHALINE CIRCULATION, ENSO, AO, AND CLIMATE VARIABILITY:

This final section discusses ENSO and the annular modes' connection to the global meridional overturning circulation (MOC), one component of which is the thermohaline circulation (THC); the other component is wind-driven. First, I will discuss sea-ice extent in the Antarctic, as possibly related to the ENSO cycle, and its effect on the MOC. Then discussion turns to the hydrologic cycle, as governed by ENSO, and its possible impact on the MOC through salinity modification. Then I will offer my own nascent hypothesis concerning ocean flow, as prompted by ENSO phases, and the resulting possible effect on salinity in the North Atlantic – home of the Atlantic sector of the MOC (known as AMOC).

Pulling this together thus far, it becomes clear, as Kerr so eloquently summarized [1], "While an ocean like the tropical Pacific may set the pace, the atmosphere can extend an ocean's reach." Could the topical Pacific be that "control central"? As this summary of oscillations has shown, fluctuations in the zonal plane extend vertically and meridionally and back again.

Coming full circle, back to how the paper began with the tropical Pacific, another look at El Nino seems appropriate. Indeed, El Nino may connect the three oceans. *James Carton and colleagues of the University of Maryland*, [96] merged fifty years of temperature and salinity observations from the oceans into a computer model. Their goal was to illuminate the evolution of a globally circumscribing thermal footprint. Emergent from the data was a wave pattern of oscillating SSTs. Warming moved from the Indian Ocean to the Pacific to the Atlantic. The nearly global journey of more than 30,000 kilometers takes about four years.

Purely atmospheric processes can partly explain the teleconnections west to east and from one basin to another; coupled atmospheric-oceanic teleconnections amplify the linkages. Ocean-flow variations, through the MOC in the global ocean, likely play a role on a variety of timescales – from decades to millennia.

Variations in strength of the AMOC have been connected to abrupt climate change in the paleoclimate record; a slowed AMOC has chilled the North Atlantic and warmed the Antarctic; a hastened AMOC has done the opposite. Since its inception – likely since the Oligocene ~ 34 mya – climate and ice cover in the northern latitudes of the Atlantic region appear to have been linked to vacillations in the magnitude of deep-water formation, and thus THC amplitude. Antarctic climate is thought to be linked in anti-phase relation to Arctic climate via this oceanic overturning connection. Proxy records of the last glacial reveal such a "see-saw" relation between the two poles. Abrupt warmings of 10°C-plus within decades in Greenland, followed by gradual cooling of comparable magnitude over centuries, oscillate with gradual millennial-scale warmings of one to three degrees Celsius in Antarctica. Warmth in Antarctica correlates with cold in Greenland and vice versa. Recent comparisons between ice-core data from Greenland and new high-resolution evidence from the Atlantic sector of the Southern Ocean region of East Antarctica show Antarctic warmings coincide with Greenland stadials (cool events) and begin significantly before Greenland interstadials (warm events) [101]. The amplitude of warming in the Southern Ocean (both the Indian Ocean and Atlantic Ocean sectors – likely well mixed via the ACC) is linearly correlated with the duration of stadials in Greenland. Implicit in this observation is that a reduction in strength of the MOC is responsible for both. A less robust AMOC allows more heat to build up in the high latitudes of the Southern Ocean, robbing it from the high latitudes of the North Atlantic[101]. A trigger for fluctuations of the MOC has long been assumed to be linked to the freshwater balance in the North Atlantic – region of deep-water formation[97]. Iceberg discharges in the North Atlantic long have been implicated in disrupting NADW formation, but timing evident in the data fails to fully support the connection. Perhaps the trigger, or an additional trigger, lies in the Southern Ocean.

Freshwater in the Antarctic may contribute to AMOC strength. Increased downwelling off Antarctica fills the Atlantic basin with Antarctic Bottom Water (AABW). This water mass competes with the NADW - deepwater formation in the North Atlantic. It follows that perhaps diminished downwelling off Antarctica might coincide with a strengthened MOC[98]. The Bolling/Allerod warming in the North Atlantic, about 14,600 years ago, roughly coincided with a meltwater pulse (leading to approximately 20 meters of sea-level rise in less than 500 years, beginning ~ 14, 200 ya). It appears that most of this melt water came from the Antarctic region from the partial collapse of the Antarctic ice sheet[99]. This would have reduced deep-water formation off Antarctica. This meltwater pulse coincided with increased NADW formation in the North Atlantic and robust AMOC flow *(Slides 39 & 40)*.

Changes in sea-ice extent (SIE) in the Antarctic may play a role in deep-water formation. More sea-ice formation would create denser ocean waters due to brine rejection upon freezing. With increased density, downwelling off Antarctica would be strengthened. Increased amounts of CO2 would be carried into the oceans, removing CO2 from the atmosphere[119]. Through competition between AABW and NADW, could this lead to slowing of the MOC? Both a slowed MOC and reduced atmospheric CO2 could lead to global cooling. Model studies suggest that Weddell Sea Bottom Water increases in response to sea-ice growth in the Atlantic sector of the Southern Ocean. This contributes to a bipolar see-saw effect - an increase in AABW and decrease in NADW [100]. Conversely, with less Antarctic SIE formation, there would be less downwelling, and perhaps a strengthened MOC.

New ice-core data allows examination on time scales of decades to centuries[101]. In light of the possible connections between ENSO and Antarctic SIE, as communicated through the coupled action of the ACW and GEW, or through atmospheric teleconnections alone[66], the possibility of the global role played by “control central” in the tropical Pacific becomes increasingly intriguing; although it is wise to hold in mind the difficulty inherent in resolving various timescales related to ENSO behavior, present and past.

Moving away from the Antarctic to a more direct path, in modern-day climate, ENSO affects the THC component of the AMOC through atmospheric processes. *Latif et al. ’99* [102] use model studies to show that atmospheric feedbacks responding to an increased frequency of El Nino events stabilize the THC through increasing salinity in the North Atlantic. The increased salinity is due to enhanced freshwater export from the Atlantic. A shift in the Walker circulation leads to anomalous subsidence over the tropical Atlantic, increasing evaporation and reducing Amazon runoff to the Atlantic. The salt anomaly that develops in the tropical region is exported poleward and mixes within the subtropical gyre. The entire North Atlantic surface salinity increases. This anomaly is carried to the region of deep-water formation where it overrides the impact of freshwater delivery due ice melt to warming in the high latitudes.

Schmittner et al. ’00 [103] base their study on a similar premise – El Nino phases result in positive salinity anomalies in the North Atlantic due to enhanced freshwater export from the Atlantic. The opposite scenario happens during a La Nina. They conclude that if the number of El Nino years is balanced by a similar number of La Nina years, the effects on the THC will balance. But, they reason, if a phase of ENSO persisted, the impact on the THC component, and therefore the AMOC, could be profound. La Nina conditions lasting longer than 70 years would lead to a collapse of the THC (or MOC) in their model studies. Prolonged periods of El Nino enhance the MOC [103-104] *(Slide 41)*.

Do teleconnections, east to west, via ocean flow, as orchestrated by ENSO phases, also play a role in modifying Earth's climate? ENSO phases regulate ocean flow through the Indonesian Throughflow [105-106, 36] and the Bering Strait [107-108]. In turn, this leads to salinity changes in the North Atlantic. During an El Nino, the warm pool is shifted eastward; the Aleutian Low strengthens and shifts southeastward. As a consequence of the former, ocean flow through the ITF decreases. Reduced flow through the ITF promotes a positive phase of the Indian Ocean Dipole (IOD) [56-58, 118]. Easterlies over the Indian Ocean strengthen. This leads to the suppression of the formation of high-salinity eddies off the southern tip of Africa via the Agulhas Current, leading to a reduction in the delivery of salinity to the Atlantic [107-111]. In addition, subtropical gyre strength in the North Atlantic weakens; slowing transport of water northward. Consequent of increased flow through the Bering Strait, fresher water from the Pacific is delivered to the Arctic region through the high-latitude route[109, 112-113]. Both ocean-flow consequences of a persistent El Nino pattern could conceivably lead to a freshening of the North Atlantic, albeit on the order of decades, leading possibly to a cooling of the North Atlantic. El Nino's relationship to a cooling North Atlantic via this mechanism, operating on the order of decades, stands in opposition to the conclusion of El Nino's more immediate enhancing effect on the North Atlantic THC through salinity increase. Whether the timing of these flow regimes coincide to create the North Atlantic freshening as envisioned by the author, is only a guess – little more than an idea on a napkin – an idea the author has nicknamed "the gateways hypothesis". But, considering these flow regimes have been evaluated individually, it seems their collective interaction and possible influence on THC flow is worthy of future study.

And finally, *Delworth and Dixon '00* [114] find that an enhanced positive phase of the AO leads to an augmentation of the THC through enhanced winds, and thus enhanced heat extraction leading to cooling and increased density. This effect persists for a decade or so, ultimately losing out to the enhanced melt water resulting from the amplified warmth in the region. If one assumes the negative phase of AO leads to the opposite result of reduced heat extraction from the ocean, and thus no augmentation of the THC, and if one considers the relationship of El Nino with a negative-phase AO, then one might speculate that this feedback works in opposition to the influence suggested by *Latif '99* and *Schmittner '00* [102-103].

Thus, on decadal timescales, with atmospherically teleconnected signals from El Nino leading to increased THC and El Nino-associated negative AO phase impacts leading to the opposite, does this imply a negating of competing feedback effects, and thus to a stabilization of the THC, at least on these timescales? And where might the ocean-gateway hypothesis fit in, if it holds any merit at all? If a persistent El Nino were to ultimately lead to a freshening of the North Atlantic over decadal timescales due to the "gateways" hypothesis, could this tip the balance in favor of a slowing of the THC? Or might an increase of SIE in the Atlantic sector of the Southern Ocean during El Nino-like regimes enhance downwelling in the Weddell Sea and lead to a further reduced MOC? Does the net balance of feedbacks reveal a relationship of a long term El Nino to a slowing THC, a cool North Atlantic, and a warm Antarctica? Or instead, do the opposing feedbacks counteract one another? And on a shorter time frame, does the interplay of ENSO-related feedbacks on the THC explain the decadal variability of strength of the THC? (*Slide 42*)

THOUGHTS IN CONCLUSION:

This paper has attempted to bring together the numerous oscillatory patterns and their interconnectedness with one another, ultimately connecting the two dominant modes of climate variability – ENSO and the annular modes. With oceanic and atmospheric signals criss-crossing the globe latitudinally, longitudinally, and vertically, feedbacks operate on a variety of timescales. The Holocene, the interglacial that began with fits and starts about 10,000 years ago, has been quite stable. It is tempting to assign this stability to these temporally diverse teleconnected feedbacks *(Slides 43 – 47)*.

If one considers the MOC to play a secondary, yet critical role in maintaining or disrupting this climate stability, and if one recognizes the multiple controls on the MOC strength, what is to be concluded about its contribution to Holocene stability? Are the tropics "control central", with ENSO phases responding to changes in heat flux and responding with signals that initiate global chain reactions that work remotely to balance the system within a narrow, oscillatory range?

Much of this paper's speculation rides on variability internal to the earth system. Ignored, to this point, is earth's major source of energy - incoming solar radiation. This paper does not pretend to offer educated insight into the influence of solar variability on climate, yet it is worth mention that its variability on decadal to orbital scales likely plays a role in behavior of both ENSO and the annular modes. Its significance is something I cannot defend or debate. But it is worth consideration, nevertheless. Increased solar variability on decadal scales, for example, during the boreal winter, weakens the Aleutian Low and shifts it northwestward [115-116]. Flow through the Bering Strait is reduced. In the tropical Pacific, the increased solar intensifies the ITCZ and SPCZ, increasing precipitation there, strengthening the SE trades, displacing the ITCZ further north, amplifying upwelling in the EEP – a La Nina-like regime.[117] The warm pool is shifted more westward, amplifying flow through the ITF. A negative phase of the IOD is induced. More saline eddies are spawned off the Agulhas Current as a result [107-108]. Increased salinity is delivered to the Atlantic. Windstress curl over the North Atlantic increases with increased solar. The Atlantic ITCZ migrates north. The subtropical gyre is strengthened. Perhaps all work together to amplify transport of heat and anomalously high salinity poleward.

In addition, increased solar variability on a decadal scale likely amplifies the Arctic Oscillation. In model studies[79] in which multiple levels of the stratosphere are allowed to participate with amplified amounts of ultra-violet radiation from the high phase of a solar cycle, stratospheric dynamics respond in such a way to strengthen the vortex. A positive AO amplifies THC flow initially, according to Delworth's hypothesis mentioned above, intensifying, for a few decades, the formation of deep water.

Could this collection of processes, nudged along by amplified solar output, work together, balancing negative and positive feedbacks to changes in the tropical Pacific and annular modes to regulate the THC flow? Would the balance be a slight net increase in the THC with increased solar output? Would this lead to slight warming in the northern high latitudes? Would reduced solar have the opposing effect?

Could this collection of processes be nudged along by other factors – episodes of intense volcanic activity, accumulation of greenhouse gases, an oscillatory behavior between a strong polar-equatorial temperature gradient and a weak one, not related to external processes? Do the positive and negative feedbacks closely balance, at least during interglacials, which have historically been extraordinarily stable, at least once the initial stages are passed?

A caveat must be mentioned here, one that may be a clue to the difference in behavior between glacials and interglacials. During a glacial and the early stages of an interglacial, when climate is highly unstable and variable, sea level is too low for flow to exist through the Bering Strait. Thus, this freshening agent would not be operative during a glacial. If flow through the BS during an El Nino could augment freshening of the North Atlantic and it was absent during a glacial, could this have led to a system prone to excessive build-up of salinity? Could this have contributed to the abrupt warmings of the Dansgaard-Oeschger events, leading subsequently, through a chain of events, to the opposite, to extreme cool events? It's an idea, obviously one with many flaws, but an intriguing possibility, noting that frequency and strength of El Nino/La Nina phases, according to *Sun et al*[23], might be a response to polar-equatorial gradients, thus forcing us to look at the entire global network as an oscillating system, with equatorial conditions forcing high-latitude conditions, which in turn, force equatorial conditions.

More questions than answers, of that there is no doubt.

GLOSSARY:

AAO	Antarctic Oscillation
AABW	Antarctic Bottom Water
AAIW	Antarctic Intermediate Water
AO	Arctic Oscillation
ACC	Antarctic Circumpolar Current
ACW	Antarctic Circumpolar Wave
CEA	Central Equatorial Atlantic
CEP	Central Equatorial Pacific
EEP	Eastern Equatorial Pacific
ENSO	El Nino Southern Oscillation
EUC	Equatorial Undercurrent
GE	Gyre Exchange
GEW	Global ENSO Wave
IOD	Indian Ocean Dipole
IOZM	Indian Ocean Zonal Mode
IPO	Interdecadal Pacific Oscillation
ITCZ	Intertropical Convergence Zone
ITF	Indonesian Throughflow
MJO	Madden-Julian Oscillation
MOC	Meridional Overturning Circulation
NADW	North Atlantic Deepwater
NAM	Northern Annular Mode
NAO	North Atlantic Oscillation
NE	Northeasterly or northeasterlies
NEC	North Equatorial Current
NECC	North Equatorial Countercurrent
NH	Northern Hemisphere
NPO	North Pacific Oscillation
OLR	Outgoing Longwave Radiation
PDO	Pacific Decadal Oscillation
PNA	Pacific North American (pattern)
PNJ	Polar Night Jet
PSA	Pacific South American (pattern)
QBO	Quasi-biennial Oscillation
QSW	Quasi-stationary waves (Rossby wavetrain)
SAIM	South Asian Monsoon
SAM	Southern Annular Mode
SE	Southeasterly or southeasterlies
SEC	South Equatorial Current
SH	Southern Hemisphere
SLP	Sea-level pressure
SO	Southern Oscillation
SOI	Southern Oscillation Index
SPCZ	South Pacific Convergence Zone
SSH	Sea-surface height
SSS	Sea-surface salinity

SST(s)	Sea-surface temperature(s)
SSW	Sudden stratospheric warmings
TAV	Tropical Atlantic Variability
TIW	Tropical Instability Waves
THC	Thermohaline Circulation
TNA	Tropical North Atlantic
WEA	Western Equatorial Atlantic
WEP	Western Equatorial Pacific
WHWP	Western Hemisphere Warm Pool
WWB	Westerly Wind Bursts
WWP	Western Warm Pool

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