Notes of the tutorial lectures for the Natural Sciences part

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First lecture

THE CLIMATE SYSTEM

"The climate is a beautiful system, exceedingly rich in interconnections and complexities."
(A. H. Oort, 1986)

WEATHER:

- **short lasting meteorological events.**
- **characteristic time scale: few days.**

CLIMATE:

- **Average state of the atmosphere (+ measures of variability) over a finite time for a certain number of years (e.g., climate of a day-night cycle, month, season, year, decade, or even longer).**
- **Statistics of weather averaged over a time period that contains many weather events, usually at least a month (e.g., the mean summer temperature, the mean February rainfall, the variance of the temperature).**
- **Interannual variation: the climatic statistics show significant variability from year to year, above the intrinsic random variability, associated with patterns that have characteristic properties in space and time. Ex.: ENSO.**

(Fig. 5, from Lau, 1992, and Schneider, 1992)

Climate and weather prediction

COMPONENTS OF THE CLIMATE SYSTEM:

- **Atmosphere**
- **Oceans**
- **Cryosphere**
- **Land/biosphere**

Atmosphere and oceans

- **Organized circulation, chaotic motions and random turbulence.**
- **React to perturbations on very different time scales.**
- **Interactions between them occur on many scales, close to their boundary.**

Atmosphere

- **Most rapidly reacting element of the system to forcing.**
- **Composition affect absorption and transmission of solar radiation**

Oceans

- **Major element in terms of long-term variability.**
- **Regulator of atmospheric temperature and gas concentrations.**
- **Storage and transport of heat and greenhouse gases.**

Land/biosphere

- **Extent, position, and orography of the continents (slowly changing)**
- **Characteristics of lakes, rivers, soil moisture, and vegetation (more rapid varying).**
- **Probably a vital climate control concerning the carbon uptake.**
- **Living organisms, particularly forests and plants, play a key role in atmospheric heat, moisture and energy budgets close to the surface.**
- **The two-way interactions between the biosphere and the atmosphere are still poorly known for their effects to be adequately included in climate modeling.**

Cryosphere

- **continental ice caps and floating sea ice - influences the surface energy balance by the high albedo and contributes to instabilities in the atmospheric general circulation as a result of temperature differences between the Poles and the Equator.**
- **It affects continental heating and upper ocean mixing and the energy exchange between the surface and atmosphere.**

HEAT CAPACITY

- **High heat capacity → low rate of temperature change in response to thermal change.**
- **heat capacity decreases with specific heat and density:**

  → the oceans respond slowly to thermal changes, and act as stabilizers in the climate system.

  → the atmosphere, respond relatively quickly to changes in the forcing function.

  →The land surface falls in between.

- **Thus, summer-winter temperature differences on land are much greater than in the oceans.**
- **Other reason for the different temperature response times of land and oceans:**

  → the mobility of the oceans

  → the partial transparency of the oceans.

INTERACTIONS

- **The climate system involves the interaction (over a wide range of differing time scales) of the air, sea, ice, land/biota, with solar radiation providing the energy that drives it.**
- **Variations of gaseous and particulate constituents of the atmosphere, along with changes in the Earth position relative to the sun, vary the amount and distribution of sunlight received.**
- **The temperature of the oceans has a marked influence on the heating and moisture content of the atmosphere.**
- **The sun radiant energy drives the atmospheric circulation, and by wind stress and heat transfer, it drives the circulation of the oceans. The atmosphere and oceans are both influenced by the extent and thickness of ice, as well as by the shape and composition of the land surface.**

(Fig. 1, first part, from Harries, 1994)

RADIATION AND ENERGY BALANCE

- **All physical things emit radiant energy in proportion to its absolute temperature, with wavelength inversely proportional to the temperature of the radiator:**

  → Sun (6000 °C): shortwave radiation (centered in the visible interval).

  → Earth (15 °C): longwave radiation (infrared).

- **Atmosphere: relatively transparent to the visible interval and a good absorber/emiter of longwave radiation.**
- **Upper atmosphere (O_3, O_2) absorbs almost all of the ultraviolet radiation**

Balance:

- **Solar energy absorbed by E/A = escaping infrared radiation**

  → effective radiation temperature of the planet: -18 °C.

- **Greenhouse effect → average surface temperature=+15 °C → agents: H_2O, CO_2, CH_4, O_3 and particles.**
- **Heat source of the lower atmosphere: the Earth's surface.**
- **Earth's energy balance closure: sensible and latent heat fluxes.**
- **2/3 of the thermal energy exchange with the atmosphere over the oceans are dominated by evaporation (latent heat) → importance of SST.**

(Fig. 1, second part, and Fig. 2, from Schneider, 1993)

Convection with release of latent heating represents a major source of energy for the atmosphere → deep cumulus clouds are used as a proxy for atmospheric heating.

HYDROLOGICAL CYCLE

The energy cycle drives the water cycle, especially the movement of water vapor in the atmosphere (latent heat flux).

Evaporation→ Condensation→ Precipitation→ Runoff.

Reservoirs:

Oceans	**95.96 %**
Ice caps and glaciers	**2.97 %**
Groundwater	**1.05 %**
Lakes	**0.009 %**
Rivers	**0.0001 %**
Soil Moisture	**0.0045 %**
Atmosphere	**0.001 %**
• **Terrestrial**	**0.0003 %**
• **Oceanic**	**0.0007 %**
Biosphere	**0.0001 %**

- **Over the oceans, evaporation exceeds precipitation: atmospheric water vapor is transported to the continents and is made up by input via runoff from the continents.**
- **Precipitation and evaporation show much variability over the continents and oceans. For precipitation: 1) sufficient water vapor in the atmosphere and 2) rising air.**
- **Net precipitation: highest near the equator (10S-10N) and in north and south middle latitudes (35-60), lowest in the subtropics north and south (15-30).**

(Fig. 3, from Berner and Berner, 1996)

HEAT BALANCE OF THE EARTH: ATMOSPHERIC/OCEANIC HEAT ENGINE.

- **More solar radiation is received per unit area at lower than at higher latitudes because:**
  - **the earth is a sphere**
  - **the duration of daylight undergoes little seasonal change in the tropics.**
- **The long-wave radiation leaving the earth varies little with the latitude.**

(Fig. 4, from Berner and Berner, 1996, and from Salby, 1993)

⇒ THERE IS AN IMBALANCE IN NET RADIATION

Why don t the poles become colder and the tropics become warmer?

Heat is transported from lower to higher latitudes by the circulation of the atmosphere and oceans by

(1) ocean currents carrying warm water,

(2) by atmospheric circulation (wind) carrying warm air, and

(3) by atmospheric circulation carrying latent heat in the form of water vapor.

The atmosphere and oceans act like a "heat engine" driven by latitudinal variations in solar radiation.

Second lecture

CLIMATE VARIABILITY

TIME SCALES - PROCESSES INVOLVED

- **Seasonal cycle (climatology)**
- **Superimposed anomalies**

→ Will be discussed mostly from a meteorological point of view due to the important role of the atmospheric circulation in climate variability and in the impacts on human systems.

→ The largest impacts come through the meteorological parameters of precipitation and temperature.

→ However, the atmospheric variability is intimately linked to the behavior of other components of the climate system.

→ We shall consider only the subset of interactions involving the atmosphere and the underlying ocean (land) surfaces.

→ The atmospheric variability in a particular time scale is often linked to a specific set of dynamical and physical processes.

→ It is important to understand the nature and origin of atmospheric variability on different time scales.

WEATHER

Synoptic time scales - Periods of several days

Fluctuations with distinct structural and propagation characteristics, that primarily owe their existence to internal atmospheric processes, such as hydrodynamic instability and convection.

TIME SCALES OF CLIMATE VARIABILITY

- **Intraseasonal variability**

  →Periods ranging from about 10 days to a season

  →relevant to medium and long-range weather forecasting.

  →linked to both the inherent atmospheric variations and alterations of the boundary conditions.

  →Ex.: persistent high pressure centers in the extratropics (blocking), intraseasonal variations with periods of 10-30 and 30-60 day (Madden-Julian Oscillation).

- **Interannual variability**

  →Periods of several years

  →mostly related to interactive processes taking place at the air-sea and air-land interfaces, in view of the long memory embedded in many maritime and land processes

  →This is the climate variability we ll be focusing on.

- **Decadal / Interdecadal variability and beyond**

  →Effects of processes with longer characteristic time scales, such as interactions with the deep ocean (thermohaline circulation) or the cryosphere, secular changes in the concentration of chemical constituents in the atmosphere, and variations in the Earth s orbital parameters.

(Fig. 5, from Lau, 1992, and Schneider, 1992)

CLASSES OF CAUSES

a) External forcing

External processes which can cause changes to the state of the climate, but which are unaffected themselves by the climate state.

Examples: changes in the orbit of the Earth, changes of solar output, continental drift and even volcanic eruptions.

b) Internal forcing

Internal interactions, instabilities and feedback processes which, even in the absence of any changes in the external forcing, can cause changes in the climatic state. Examples: El Ni o, changes in the production or loss of greenhouse gases.

What is internal on long-time scales may be external on shorter ones, depending on what processes are included in the climatic system defined for an investigation.

For instance, the atmosphere is forced by SSTs that themselves respond gradually to atmospheric forcing, in a complicated feedback. Therefore, for shorter periods, one can simply use the SST as a prescribed forcing to the atmosphere, and in this case, it would be an external forcing for the atmosphere.

THE ATMOSPHERIC CIRCULATION

Equations of motion

a) The momentum equation

$$\frac{D\vec{V}}{Dt} = -2\vec{\Omega} \times \vec{V} - \frac{1}{\rho}\vec{\nabla} p - g\vec{k} + \vec{F}$$

The horizontal component is:

$$\frac{D\vec{V}_h}{Dt} = -f\vec{k} \times \vec{V}_h - \frac{1}{\rho} - \vec{\nabla}_h p + \vec{F}$$

where $f = 2\Omega sen\phi$ is the Coriolis parameter.

The driving force of the atmosphere is the pressure gradient force, but the direction of the motion is changed by the Coriolis force, an apparent force induced by the rotation of the earth, that deflects moving air parcels to the right in the Northern Hemisphere, and to the left in the Southern Hemisphere.

The vertical component is approximated by the hydrostatic equilibrium:

$$\frac{\partial p}{\partial z} = -\rho g$$

b) The continuity equation

$$\frac{D\rho}{Dt} + \rho(\nabla.\vec{V}_h + \frac{\partial w}{\partial z}) = 0$$

c) The thermodynamic equation:

$$c_v \frac{DT}{Dt} = -p\frac{D\alpha}{Dt} + \dot{Q}$$

Complementary conservation equations of some atmospheric constituents: $\frac{Dq_i}{Dt} = S_i$

In these equations: $\frac{D}{Dt} = \frac{\partial}{\partial t} + \vec{V}.\nabla$ is the material time derivative.

Circulation Features

- **Near the equator**

  The Earth's rotation has a smaller influence on air motions. Kinetic energy is associated with "thermally direct circulations" forced by distribution of atmospheric heating (pressure gradient)

- **Away from the equator**

  In the higher troposphere (low friction) the pressure gradient force is nearly balanced by the Coriolis force (geostrophic equilibrium):

  $$f\vec{k} \times \vec{V} = -\frac{1}{\rho}\vec{\nabla}_h p \quad \Rightarrow \quad \vec{V}_g = \frac{1}{\rho f}\vec{k} \times \vec{\nabla}_h p$$

  The wind blows almost parallel to the isobars (or to the contours of geopotential height) and its magnitude increases with the pressure gradient (ou geopotential height gradient).

  Pure geostrophic motions are non-divergent. Thus, there is no vertical motion. Vertical motion is produced by mechanisms that drive the horizontal velocity ageostrophic and introduce divergence (e.g., friction).

- **Atmospheric waves**

  The most important type for climate variability are the long Rossby waves.

  → Origin: the variation with latitude of the Coriolis parameter provides a torque on air parcels moving meridionally.

  → Large spatial scales, 1,000 km and greater (planetary waves), and characteristic time scales of days and longer.

  → The smaller scale synoptic waves are also Rossby waves, but they originate in instability of the zonal circulation.

  → Planetary waves are excited by orography and by latent heat release in organized convection.

(Fig. 6, 7, 8, from Salby, 1993)

THE GENERAL CIRCULATION

a) Zonally symmetric picture

- Origin: the latitudinal imbalance of heating creates difference of pressure which drives a meridional overturning of air, with rising motion at low latitudes and sinking motion at subtropical and high latitudes.
- If the circulation were due solely to heating ⇒ two symmetric closed meridional cells (SH and NH).
- Other factors (especially the Earth s rotation) ⇒ three pairs of cells and a prevalent zonal circulation.

b) Picture with zonal asymmetries

Differential heating is not only caused by latitudinal variations of solar input ⇒ zonal asymmetries in the heating, due to the relative distribution of latent heating in the tropics, land and ocean, effects of vegetation and land use, surface topography and roughness and so on.

Zonal asymmetries in the distribution of sea and land ⇒ the subtropical high-pressure belt is broken into low-level centers of high pressure over the oceans and low pressure over the continents in summer. Monsoon circulation.

The circulation is such that the subsidence is stronger in the eastern part of these highs (coastal deserts of Northern Chile and Lower California). In the western region, convergence and ascent appear to be more prevalent (SPCZ, SACZ).

Zonal asymmetries in the release of latent heat in the tropics ⇒ Walker circulation.

Extratropics, in the higher troposphere ⇒

geostrophic approximation ⇒ zonal westerlies.

- **The strongest westerlies concentrate in a relatively narrow band of greatest temperature (pressure) gradient. The wind speed increases with the altitude, because the pressure gradient increases with the altitude ⇒ jet stream.**

- **There is only a small mean meridional component, but the mean consists of many transient patterns that transfer heat between the equator and the poles.**

- **The westerlies are subject to instability because of the earth's rotation. This introduces waves in the instantaneous circulation (cyclonic-anticyclonic systems), that deflect air meridionally, transferring energy, momentum and water vapor between low latitudes and high latitudes and from the surface to the top of the atmosphere.**

- **Surface cyclones develop to the east of the throughs of these waves, and produce a series of storms that travel from west to east around the globe (synoptic weather systems).**

- **Storm tracks: preferred trajectories of these disturbances, located in the extratropical oceans. During the winter season, downstream and slightly poleward of the quasi-stationary jet streams. Regions of strong heat and momentum transports, resulting in enhanced energy exchanges between the transient disturbances and the more slowly varying basic flow.**

- **The zonal symmetry of the westerlies is disturbed by mountain ranges (Himalayas, Alps, Rockies, Andes) and latent heating, as in the ITCZ or monsoon heating, or anomalous heating such as that associated with an El Ni o event.**

(Fig. 9, from Salby, 1993; Fig. 10, from Harries, 1994; Figs. 11 to 16)

THE OCEANIC CIRCULATION

Basic concepts

→ Oceans: 71% of the Earth s surface and Earth s principal time varying reservoirs of thermal energy and moisture.

→ Difference of temperature atmosphere/ocean ⇒ exchange of heat and moisture between the two fluids.

→ The temperature of the atmosphere adjusts to the ocean temperature ⇒ climate variations over a large part of the Earth, are strongly influenced by the changing sea surface temperature (SST).

→ The oceanic circulation is wind-driven in the surface layers and thermally driven in the deep waters.

→ Ocean circulation transports thermal energy from where there is a heat gain from the atmosphere near the surface (in the tropical and subtropical areas) to where there is a transfer of heat back to the atmosphere (in subpolar and polar latitudes).

→ The water cooled in subpolar and polar latitudes, sinks by thermal convection. Evaporative cooling and increase of salinity plays an important role there.

→ In the tropical and subtropical latitudes water rises principally where winds force a surface divergence.

→ Variations in the ocean temperature can occur over the entire water depth when net surface heat flux distributions are disturbed or when the ocean circulation changes.

→ The equations that describe the motions of the oceans are essentially the same as for the atmosphere.

→ For circulation studies, two layers: the surface layer or mixed layer (~50-300 m), and the deep water. They are separated by the thermocline, a region of steeply decreasing temperature (~1000m), with limited comunication between the surface and the deep water.

(Fig. 17, from Harries, 1994)

Shallow circulation (wind-driven)

→ Wind-stresses drive the circulation of the surface ocean, but the Coriolis force and friction are also involved, and the water is not simply carried downwind.

→ The circulation pattern can be summarized as a number of current rings, or gyres, that flow clockwise in the NH and counterclockwise in the SH.

→ Each gyre has a strong poleward component on the western side (e.g., the Gulf Stream), and weaker current on the east.

→ The wind stresses on the ocean force the surface layer of water together in the subtropics, and pull it apart in the subpolar areas and on the equator (due to the changing in sign of the Coriolis parameter). Therefore, we must have a downward vertical motion at the base of the mixed layer in the subtropics and an upward vertical motion in the subpolar areas and on the equator.

→ Equatorial upwelling causes the low SSTs in the eastern Atlantic and eastern Pacific. Major upwelling occurs along the tropical western boundary of the continents (eastern part of the ocean) where the surface currents flowing toward the Equator are broad and relatively weak.

→ Coastal upwelling: winds blowing equatorward along the coasts are deflected offshore bring about a transport of deeper water. This deeper water is enriched in nutrients, which results in high planktonic productivity and teeming life, including abundant fish (e.g., Peru and Chile and the California coast).

→ Upwelling also results when surface waters are blown away from an open-ocean area, bringing about divergence. Subsurface waters move in to replace the missing surface water. In the sea around Antarctica there are such upwelling areas.

(Fig. 18, from Niiler, 1993, and Figs. 19 and 20)

Deep circulation (thermohaline)

→ Driven by density differences arising from differences in temperature and salinity. The density stratification is due primarily to the decrease of temperature with depth. The stratification severely inhibits vertical motion⇒ the deep water circulation can be viewed as primarily horizontal.

→ The density differences between water masses of the deep sea owe their origin to surface processes. Surface water migrating to high latitudes becomes more dense because of evaporation, loss of sensible heat and sea-ice formation. In certain locations in the far north and far South Atlantic, the cool, salty surface water occasionally becomes denser than the underlying water and sinks.

→ Once it reaches great depths, the water tends to conserve its temperature and salinity as it flows laterally throughout the oceans, bringing about the deep-water circulation.

→ The lateral deep water (thermohaline) circulation: North Atlantic Deep Water flows away southward from its source on the western side of the North Atlantic, meets a strong northward-flowing current of Antarctic Bottom Water in the South Atlantic, and they merge and flow through the Antarctic into the deep Indian Ocean and ultimately into the deep Pacific Ocean. Thus, the deep water originates only in two areas, and both are in the Atlantic Ocean.

→ There is considerable interest in the thermohaline circulation of the North Atlantic and the rate of formation of the North Atlantic Deep Water because of its climatic implications. Its variations are thought to be connected to climate variability in longer time scales.

(Fig. 21 from Berner and Berner, 1996, and Fig. 22)

Third lecture

CLIMATE VARIABILITY IN INTERANNNUAL TIME SCALES

PROCESSES INVOLVED

- **Air-sea-interactions in the tropics**
- **Tropical-extratropical interactions**
- **Air-sea interactions in the extratropics**
- **Air-land interactions**

AIR-SEA INTERACTIONS IN THE TROPICS

a) Introduction

- **At large scales each medium is strongly controlled by the other and thus a large fraction of tropical climate variability may be attributed to air-sea interactions:**
    - **the large-scale upper ocean circulation is largely determined by wind stress**
    - **the major features of the tropical atmospheric circulation, averaged over timescales longer than a month or two, are largely determined by SST. The tropical SST determine the locations of regions of persistent precipitation.**
- **The internal variability of the tropical atmosphere shows sufficiently short time-scales so that a separation can be made between atmospheric internal variability and that associated with slower evolution communicated by SST.**
- **A given wind change generates stronger response at the equator than at higher latitudes; equatorial waves are less susceptible to the destructive influences of friction and mean currents.**

(Fig. 10, from Harries 1994; Figs. 11, 12; Fig. 18, from Niiler 1993; Figs. 19, 20; Fig. 23, from Harries 1994)

b) Pacific & ENSO

Climatology of the Tropical Pacific

→ The SSTs drive direct thermal circulation cells in this region: the Hadley cell and the Walker circulation. Both contribute an easterly component to the surface winds.

→ Convection tends to organize over the warmest SST, producing the ITCZ, and the strong convection over the Maritime Continent.

→ The westward wind stress is balanced by pressure gradient associated with a sea level gradient of about 40cm and a corresponding slope in the thermocline.

→ The westward surface currents and deflection due to Coriolis force to either side of the equator drives a narrow band of upwelling in the eastern/central Pacific.

→ The upwelling and shallow thermocline produces the equatorial cold tongue in the east, while the deep thermocline in the west is associated with the western Pacific warm pool .

→ The low SSTs in the cold tongue region are caused by wind-driven ocean dynamics rather than by differences in surface heating.

The ENSO cycle

ENSO arises as a self-sustained cycle: an initial SST anomaly in the eastern Pacific produces wind anomalies that enhance the inicial SST perturbation in a positive feedback, which can lead to instability.

However, what are the mechanisms for the turnabout from the state with warm SST anomalies through a state with little SST signature, to the subsequent cold phase?

The explanation uses some features of the equatorial ocean dynamics that governs the variations in the upper tropical ocean over time scales of a few years or so. For the time and space scales relevant to El Nino only two types of wave motions matter: long Rossby waves and equatorial Kelvin waves.

→ Kelvin waves: only in the equatorial region and propagate eastward at least three times faster than the Rossby waves. Caused by the wanishing of the Coriolis force on the equator.

→ Long Rossby waves: propagate energy westward

→ A Kelvin wave reflects into Rossby waves at an eastern boundary and Rossby waves reflect into a Kelvin wave at a western boundary.

Initially: warm SST anomaly in the eastern Pacific (deeper thermocline than normal).

Westerly wind anomaly, forcing a Kelvin wave packet in the ocean that further depress the thermocline in the east.

When the Kelvin wave hits the eastern boundary, it reflects as a sum of Rossby waves: some act to extend the equatorial wave-guide up and down the eastern boundary and those at low latitudes carry much of the mass and energy brought east by the Kelvin waves back toward the west.

The excess of warm water in the east must be compensated by a region of colder water (shallower thermocline). This is made in the form of equatorial Rossby wave packets, which propagate westward. At the western boundary (much later than the Kelvin waves reach the eastern boundary) they are reflected as cold equatorial Kelvin waves, which propagate eastward across the ocean to reduce the SST there. Thus the original warm signal is accompanied by a cold signal - but with a delay → delayed oscillator.

ENSO irregularity : a possible chaotic behavior, perhaps arising from the interaction of the slow components of the ocean-atmosphere system with the seasonal cycle, or stochastic forcing by uncoupled atmospheric variability (weather noise).

b) Atlantic

- **The annual movement of the ITCZ and the amount of seasonal rainfall in the underlying semi-arid regions, display coherence with the sea surface temperatures gradient across the equator.**
- **The strength of this gradient varies inter-annually due to large-scale SST anomalies that form on both sides of the equator.**
- **These tropical sea-surface temperature anomalies may vary out of phase with one another in a quasi periodic manner of 10-13 years, forming what is known as the Tropical Atlantic Dipole .**
- **Impacts of the variability of the north south SST gradient: rainfall in Northeastern Brazil, distribution and intensity of hurricanes in Gulf of Mexico and North America seaboard.**
- **Is the dipole really a mode of variability? On interannual time scales the SST variability across the equator is not correlated, but the dipole seems stronger on interdecadal time scales.**
- **On the interannual time scales there are ENSO-induced Atlantic SST fluctuations that lag their Pacific counterparts by 4-5 months and this probably disrupts the dipole. The ENSO influence is transferred through the atmospheric perturbations (for instance, the Walker cell, PNA, etc.).**

(Fig. 24, from PACS Implementation Plan)

Two hypotheses to explain the variability of the cross-equatorial SST gradient:

- **it stems from regional ocean-atmosphere positive feedbacks involving primarily SST and wind-induced latent-heat flux that acts to enhance SST variability both north and south of the equator in the tropical Atlantic Ocean, while ocean processes set the slow time scale of variability.**
- **The other hypothesis says that the SST anomalies on either side of the equator are dynamically independent and controlled by processes in each hemisphere Therefore, the variability of the cross-equatorial SST gradient would be largely stochastic in nature.**

<u>Fourth lecture</u>

TROPICAL-EXTRATROPICAL INTERACTIONS

Anomalous tropical SST

$\Downarrow$

Anomalous convection

$\Downarrow$

Anomalous latent heat source

$\Downarrow$

Anomalous ascending motion over the source (divergence)
and subsidence (convergence)
in local Hadley cell and modified Walker circulation.

$\Downarrow$

ROSSBY WAVE SOURCE!

(Fig. 25)

FACTORS AFFECTING THE EXTRATROPICAL RESPONSE:

- **Interactions with the mean flow (zonal asymmetries $\rightarrow$ secondary sources)**
- **Feedback from transients in storm tracks (Rossby wave patterns $\rightarrow$ affect storm track $\rightarrow$ some do it in a way as to produce positive feedback) $\Rightarrow$**
- **Preferred modes .**

TELECONNECTION PATTERNS

- Patterns that describe simultaneous variations of atmospheric circulation (or other parameters).
- A physical reason for the simultaneous variations is implied.
- Examples and maps (PNA, PSA).

SEASONAL DEPENDENCE

During ENSO, the extratropical response:

- is stronger in the NH winter (stronger westerlies and closer to the equator, and larger SST anomalies)
- is significant in the SH in all seasons, but the impact over SSA is stronger in spring (still strong westerlies in the subtropics and already large SST anomalies).
- shows regional impacts in summer.

REGIONAL IMPACTS ON CLIMATE

PNA: warmer along the west coast of NA
southward shift of the storm track
cold breaks accross the plains of NA

PSA: more rainfall over Chile

SOURCE REGIONS FOR TELECONNECTION PATTERNS

Motivation:

Dynamical links → better forecast because:

- **better statistical models**
- **better assessment of dynamical models**

Influence functions → indicate the regions where the anomalous heating is most effective in producing circulation anomalies at a given point.

Examples - (Figures in Grimm and Silva Dias, 1995)

AIR-SEA INTERACTIONS IN THE EXTRATROPICS

In the tropics→ **SST changes are cause of large-scale atmospheric anomalies.**

In the extratropics→SST changes are primarily response to, rather than causes of atmospheric changes

(Fig. 14 of Trenberth et al. 1998)

Influence of tropical X extratropical SST

Results of models:

- tropical Pacific SST produces stronger atmospheric response than SST in the extratropics
- atmospheric heating effect may not be local in the extratropics
- atmospheric bridge between SST variability in the tropics and extratropics
- SST patterns: oceanic response to atmospheric driving

positive feedback from the SST to the atmosphere

(Fig. 12 of Trenberth et al. 1998)

Examples of Air-sea interactions in the extra-tropics

North Pacific

SST in ENSO region:

- **positive correlation with SST of the western seabord of North America**
- **negative correlation with SST in Central North Pacific**

Cause: PNA pattern → cyclonic circulation in NP:

→ cold and dry advection to the west of the low → cold SST anomalies in central NP → warm and moist advection to the east.

→ storminess to the south.

South Pacific

Something similar (?)

INTERDECADAL VARIABILITY AND AIR-SEA INTERACTIONS IN THE EXTRATROPICS

a) Pacific Decadal Oscillation (PDO)

- **Interaction with ENSO**
- **Periods above 20 years**
- **also associated with PNA→impacts**
- **What is the dynamics involved in the decadal oscillations?**

(Figures in: www.pmel.noaa.gov/~miletta/web/pdo_p1.html
and:
tao.atmos.washington.edu/PNWimpacts/REPORTS/pdo_clim_pics.html)

b) North Atlantic Oscillation (NAO)

- **Interannual and interdecadal variability**
- **seesaw of anomalous pressure between Iceland Low and Azores High**
- **Links with SST and oceanic currents**
- **Links with subtropical SST (dipole?)**
- **What drives this pattern? Variations in the thermohaline circulation?**
- **Atmospheric bridge between tropical SST and extratropical circulation (and SST)?**

(Figures in: geoid.mit.edu/accp/avehtml.html)

AIR-LAND INTERACTIONS

Air land interactions compared to air-sea interactions:

- The thermal storage is less in the land.
- There is no overturning or circulatory motions to redistribute energy globally;
- Since land is not always moist, the role of latent heat transfer is less extreme than for the oceans.

On the other hand:

- land is more variable and changeable than the oceans for many of the important coupling processes:
- when wet, it can exchange water with the atmosphere more rapidly than the oceans because of greater surface roughness, but when dry it provides no water at all;
- local temperature are much more responsive to net radiation than are the oceans. The annual temperature range over the oceans is much smaller than over land.
- Presence or absence of clouds has a substantial effect;
- the albedo varies with type of surface cover, and this cover is highly variable in space (and even in time).

Different boundary conditions on a mesoscale can alter mesoscale wind and precipitation patterns. Only massive changes would aggregate to significant changes on a continental or larger scale. However, as they may be significant in the regional scale, they must be taken into account. During El Ni os or La Ni as, for example, changes in land surface hydrology and moisture availability can feedback and influence the total response.

Different boundary conditions over land may also affect, for example, the interannual variability of monsoon. Examples.

(Fig. 22.3 and 22.6, from Dickinson, 1993)

Fifth lecture

PREDICTABILITY

1. Weather predictability

The atmosphere is a chaotic system: very sensitive to changes in its initial conditions.

(Figure representing stable and unstable systems)

Limit of deterministic predictability: given by the rate of grow of the inevitable errors in the initial state.

Lorenz (1982): the precise state of the atmosphere cannot be predicted more than 2 weeks in advance.

2. Why is climate predictable?

Climate is concerned with the statistics of the atmosphere over a given period (e.g., monthly or seasonal mean and variance of precipitation and temperature).

The statistics of the atmosphere depend on the boundary conditions (temperature, reflectivity, surface moisture,).

The key to climate prediction is predicting the boundary conditions.

3. Predictability of SST

The irregularity of ENSO may be due to stochastic forcing by weather noise ⇒ this may imply a fundamental limit to predictability (?)

Problems of

- **initialization, quality and density of data**
- **limitations of the models.**

Skill - example of the Cane-Zebiak model.

(Figure of Chen et al. 1997)

4. Predictability of tropical climate

Tropical atmospheric circulation:

- **driven directly by the latent heat release in regions of persistent precipitation (determined by SST).**
- **Latent heating → rising motion → upper level divergence →lowered surface pressure → low level convergence → ...**
- **Dominated by the Hadley and Walker cells.**

In the tropical region:

- **SST translates into temperatures over land and precipitation**
- **There is weaker non-linear coupling between different time scales than in the extratropics.**

5. Predictability of extratropical climate

Tropical SST anomalies shift regions of persistent precipitation → circulation anomalies propagate to the extratropics (Rossby waves).

The extratropical weather variability is high, driven by dynamically unstable synoptic systems → chaotic nature.

However, there is considerable impact of the low-frequency large-scale anomalies generated in the tropics (for strong SST anomalies).

The signal is embedded within the noise of natural variability.

Noise → unpredictable internal variability

Signal → predictable, externally forced variability.

Signal / Noise → measure of predictability

- dependent on the season
- dependent on the time averaging.

Other analysis tool: PDF of atmospheric states (2nd moment).

(Fig. 13 of Trenberth et al. 1998).

Sixth lecture

CLIMATE PREDICTION TECHNIQUES

DESCRIPTION

CAPABILITIES

LIMITATIONS

1. DYNAMIC MODELS

Primitive equations + parameterizations

Ensembles

- **ensemble mean**
- **uncertainty**

Validation

2. STATISTICAL TECHNIQUES

Depend on correlations between the used predictors and the expected predictands.

Based on existing data

Choice of predictors.

Seventh lecture

DOWNSCALING

1. MOTIVATION

- **Computational resources limit the spatial detail a global GCM can resolve;**
- **The climatic impacts community and policy makers often need information at much finer scales;**
- **Processes acting on local or regional scales often feed into the climate system;**
- **Some specific applications need data at finer resolution.**

2. METHODS

- **Statistical techniques**

- **Coupling of mesoscale or limited area models to GCMs.**
    - Results from a GCM are used as initial and boundary conditions for a LAM, which operates at much higher resolution and frequently with more detailed physical parameterizations.

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