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"Biennial Oscillation of Rainfall in Asia/Australian Monsoon Region"

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2.1 Biennial oscillation of rainfall in Asia/Australian monsoon region

The interannual variability of monsoon rainfall over India and Indonesia/ Australian region show a remarkable biennial rhythm. The spectral analysis of the long-term homogeneous record of monsoon rainfall averaged over the whole of India has shown a dominant peak of 2.2-3 year period, i.e., quasi-biennial oscillation (QBO) (Mooley and Parthasarathy, 1984). This periodicity also have appeared as a dominant periodicity in the rainfall series in Indonesia (Yasunari and Suppiah, 1988) and in east Asia (Tian and Yasunari, 1992; Shen and Lau, 1995). Thus, the BO or QBO-like oscillation seems to be a fundament nature of Asian/ Australian monsoon rainfall. In reality, this QBO turns out to be a flucutation that the anomay of monsoon rainfall in one year tends to be reversed in sign in the next and/or the previous year.In this sense, this QBO has a different characteristic from that in the zonal wind of the equatorial stratosphere (Yasunari, 1989).

2.2 Spatial tructure of QBO in monsoon and atmosphere/ocena system

The QBO in the rainfall in Asia/Australian region manifest itself as part of the QBO in the monsoon and the atmosphere/ocean system of this region, which has a charasteristic spatial structure and seasonality as well. Meehl (1987), by stratifying anomalies of oceanic and atmospheric parameters of each successive seasons in terms of strong (and weak) Indian monsoon, noticed that the anomalous state of the atmosphere/ocean system in the Asian/Australian monsoon region starts from northern summer and evolve toward the following southern summer as shown in Fig. 1(Meehl, 1993). The anomalies in covection and SST moves its center from northwest toward southeast over this region, following the seasonal march of the system.

The lower-tropospheric wind field associated with this QBO in SST shows an out-of-phase relation beween the Indian Ocean and the Pacific Ocean basins as shown in Fig. 2 (Ropelewski and Halpert, 1992), with an eastward phase propagation of these two fields from the Indian Ocean side toward the Pacific Ocean (Yasunari, 1985; Kutsuwada, 1988; Rasmusson et al., 1990; Shen and Lau, 1995). This suggests that the QBO in the monsoon rainfall over india and the atmosphere/ocean system in the equatorial Indian through Pacific Ocean basin is tightly linked each other as a coupled monsoon/atmosphere-ocean system (as referred to as MAOS by Yasunari and Seki, 1992). Another feature to be noted in the near-surface wind field is anomalous cyclonic (or anti-cyclonic) circulation over the subtropical western Pacific and the South China Sea region (Rasmusson et al., 1990; Tomita and Yasunari, 1993; 1996), which is extremely significant in the QBO-scale ENSO, but is not so in the longer-time scale ENSO cycle (Tomita and Yasunari, 1993). This feature may be related to an interaction between the tropics and the extra-tropics as discussed later.

2.3 Seasonality of QBO and the "monsoon year"

It is intersting to note that while these anomalies move and develop continuously from the northern summer to the following northern winter (southern summer) season they are ceased and decayed beween the northern winter and the following northern summer season. This asymmetric relation between the monsoon rainfall and the atmosphere/ocean system in the seasonal sequencee is more directly shown in the time series of Indian monsoon rainfall anomaly and the surface sea water temperature of the equatorial western Pacific in the subsequent winter (January) as shown Fig. 3 (Yasunari, 1990). A more systematic relation between the monsoon activity in the northern summer and the atmosphere/ocean system is shown in the lag-correlation beween the Indian monsoon rainfall and SST in the equatorial western and eastern Pacific (Fig. 4 (Yasunari, 1990)). A remarkable feature in this diagram is that the correlation gradually increases after the summer monsoon season and reaches its maximum in the following northern winter (January) of Y(0)/Y(+1) both in the western and the eastern Pacific with the opposite signs between the two. A very similar feature of lag-correlation is also noticed for the east Asian monsoon rainfall (Shen and Lau, 1995). This relationship between the summer monsoon rainfall in Asia and the SST in the equatorial Pacific strongly suggests that the Asian summer plays a significant role in forming the anomalous state of the atmosphere/ocean system over this region. That is, a strong (weak) summer monsoon tends to lead La Nina (El Nino) condition in the equatorial Pacific in the later seasons of the year.

Another significant feature of this diagram is a near-zero correlation in the boreal spring to early summer monsoon season of each year, when a change of sign of correlation occurrs before and after this season. This remarkable persistency and tendency of the lag-correlation in the seasonal cycle is also apparent in the auto-correlation of the indices of the atmosphere/ocean system (e.g., SST, SOI etc.) as noted by Yasunari (1990) and Webster and Yang (1992). This implies that an anomalous state of the atmosphere/ocean system in the equatorial Pacific ocean basin tends to decay in this particular season and the other state with opposite sign tends to develop associated with starting the next summer monsoon. Webster and Yang (1992) referred to this prominent feature of boreal spring as "predictability barrier" of climate systém in the tropics.

In other words, the biennial oscillation in the ENSO-monsoon system, or the MAOS is an oscillation which tends to have a strong seasonality with the maximum-amplitude phase in boreal winter and a node phase in boreal spring to early summer. Yasunari (1991) defined this unit year starting from a boreal spring and ending at the next boreal spring as a "monsoon year" for the interannual climate variability in this region. This "monsoon year" concept can be adopted not only for the Asian/Australian monsoon region but also for other regions in the tropics, as is shown in the seasonality of anomalous rainfall in various regions in the tropics (Fig. 5 (Yasunari, 1991)).

2.4 Possible mechanisms of QBO in the MAOS

2.4.1 Atmosphere-ocean interaction in the warm water pool

Mechanisms for the QBO in the MAOS in the Asian/Pacific sector proposed so far may be classified into two groups. One is basically attributed to some feedbacks in the seasonal cycle of the atmosphere/ocean interaction in the warm water pool rgion. Nichlls (1978) noted the effect of seasonal change of feedback between wind field and surface pressure. In monsoon westerly season, wind speed anomaly is negatively correlated to prssure anomaly, while in easterly (dry) season it is positively correlated to pressure anomaly. Wind speed anomaly is, on the other hand, negatively correlated to SST change through the year, through the physical processes of evaporation and mixing of the surface ocean layer. He noted that a simple combination of these two feedbacks in the course of seasonal cycle induces a biennieal oscillation of these anomalous fields. Meehl (1993) further substantiated this idea, but in his case, focusing on the memory effect of oceanic mixed layer. That is , when large-scale convection over the warm water pool region, associated with seasonal migration of ITCZ and monsoon, is stronger (weaker), the ocean surface temperature there will finally become aomalously low (high) through the processes mentioned above. The anomalous state of SST thus produced may be maintained through the following dry seasons and even to the next wet (monsoon) season, which produces weaker (stronger) convection. The seasonal evolution of SST and convection anomalies over this region observed by Meehl (1987) seems to be , to some extent, consistent with this idea. In these hypotheses, the atmosphere-ocean interaction over the warm water pool is of paramount importance, some but elemental aspects of interannual variability of the MAOS, e.g., asymmetirc seasonal evolution of anomalies from boreal summer to winter and winter to spring as shown in Fig. 4, are not specifically explained. In addition, some phenomenological asptects outside of this region, such as snow cover - monsoon relation (i.e., Hahn and Shukla, 1976) may be treated as a passive indicater of variability of this system.

Gray et al. (1992) proposed a hyopothesis that the stratospheric QBO influences the ENSO/monsoon system in the tropics, via the vertical zonal wind shear change in the upper troposphere associated with the QBO, which in turn modulates convective activity in the equatorial and sub-tropical western Pacific. Though this is an interesting and unique idea, further examination may be needed.

2.4.2 Global-scale land-atmosphere-ocean interaction

The other QBO mechanism is attributed to more global land-atmosphere-ocean interaction. As has been pointed out in many literatures, the anomlous land surface conditions, e.g., snow cover and soil moisture, are likely to influence the seasonal land/ocean differntial heating, and thus produced anomalous Asian summer monsoon (Hahn and Shukla, 1976; Dickson, 1984 etc.). The modelling studies (Barnett et al., 1989; Yasunari et al., 1991) demonstrated that two physical processes, albedo effect and hydrological effect, may be responsible for the anomalous seasonal heating of the atmosphere. A schematic diagram of these two effects are shown in Fig. 6 (Yasunari et al., 1991). Barnett et al. (1989) suggested that the anomalous excessive snow cover over Eurasia may trigger ENSO-like condition in the equatorial Pacific by means of weakened summer monsoon and intensified equatorial westerlies, though this explains only one way process from the land-atmosphere interaction to the atmosphere/ocean interaction in the seasonal cycle.

Yasunari and Seki (1992) found significant time-lag relations between Eurasian snow cover in boreal spring (April), Indian summer monsoon rainfall, and surface water temperature in the equatorial western Pacific in the following winter. The seasonal hemispheric circulation anomalies composited for strong (weak) Indian monsoon years also strongly suggested the stationary wave patterns (Fig. 7) forced by the anomalous MAOS state in autumn to winter, which in turn is likely to be responsibe for more (less) extensive snow cover anomaly over Eurasia leading to the weak (strong) monsoon for the following summer. This may be an evidence for the global-scale QBO mechanism, though a clue may be how robust the circulation anomalies evolves for the strong (or weak) monsoon years, irrespective of chaotic nature of mid-latitude westerlies.

Meehl (1994) demonstrated in a experiment of 70 years integration of coupled ocean-atmosphere GCM experiment a very similar biennial cycle of the monsoon-atmosphere/ocean system in concert with the mid-latitude flow regimes. In this BO cycle, a combined effect of anomalous convection over the Asian monsoon and African monsoon is stressed for producing the circulation anomalies in the mid-latitudes as schematically shown in Fig. 8. This experiment also suggested that cold air advection (and cooled suface) prevailing over the Eurasian continent in winter may be more important than the effect of anomalous snow cover itself for producing anomalous monsoon and convection in later seasons.

Tomita and Yasunari (1996) have recently noted a role of east Asian winter monsoon played in the global-scale QBO. They notice a strong negative correlation between the Indian summer monsoon and the following SST anomaly in the South China Sea, which is indicative of strength of the east Asian winter monsoon surge.

That is, a stronger south Asian summer monsoon is likely to produce a stronger east Asian winter monsoon and vice versa, presumably through atmospheric teleconnection (Yasunari and Seki, 1992) and associated land surface processes (e.g., snow cover) over the Eurasian continent. The SST anomaly in the South China Sea thus produced in winter persist through the following spring and early summer, which may be very important for initiating the reversed anomalies of convective activity and monsoon in the equatorial western Pacific. The wind-stress curl mechanism assoicated with the winter monsoon surge (Masumoto and Yamagata, 1991) reinforces to change SST anomaly over the western Pacific from winter to the following summer. The overall processes of this QBO is illustrated in Fig. 9.

These global-scale QBO mechanism stresses an essential role of the two-way tropical-extratropical interaction in the seasonal cycle, in addtition to the atmosphere-ocean interaction over the equatorial Oceans. The land-surface processes over the Eurasian continent, in this idea, may be part of this interaction. In this sense, this global-scale QBO mechanism may provide a more generalized idea of the QBO, which may be able to interpret irregularities and modulation of this oscillation due to the chaotic behavior of the mid-latitude westerly flow and longer-time change of the basic state of the climate system.

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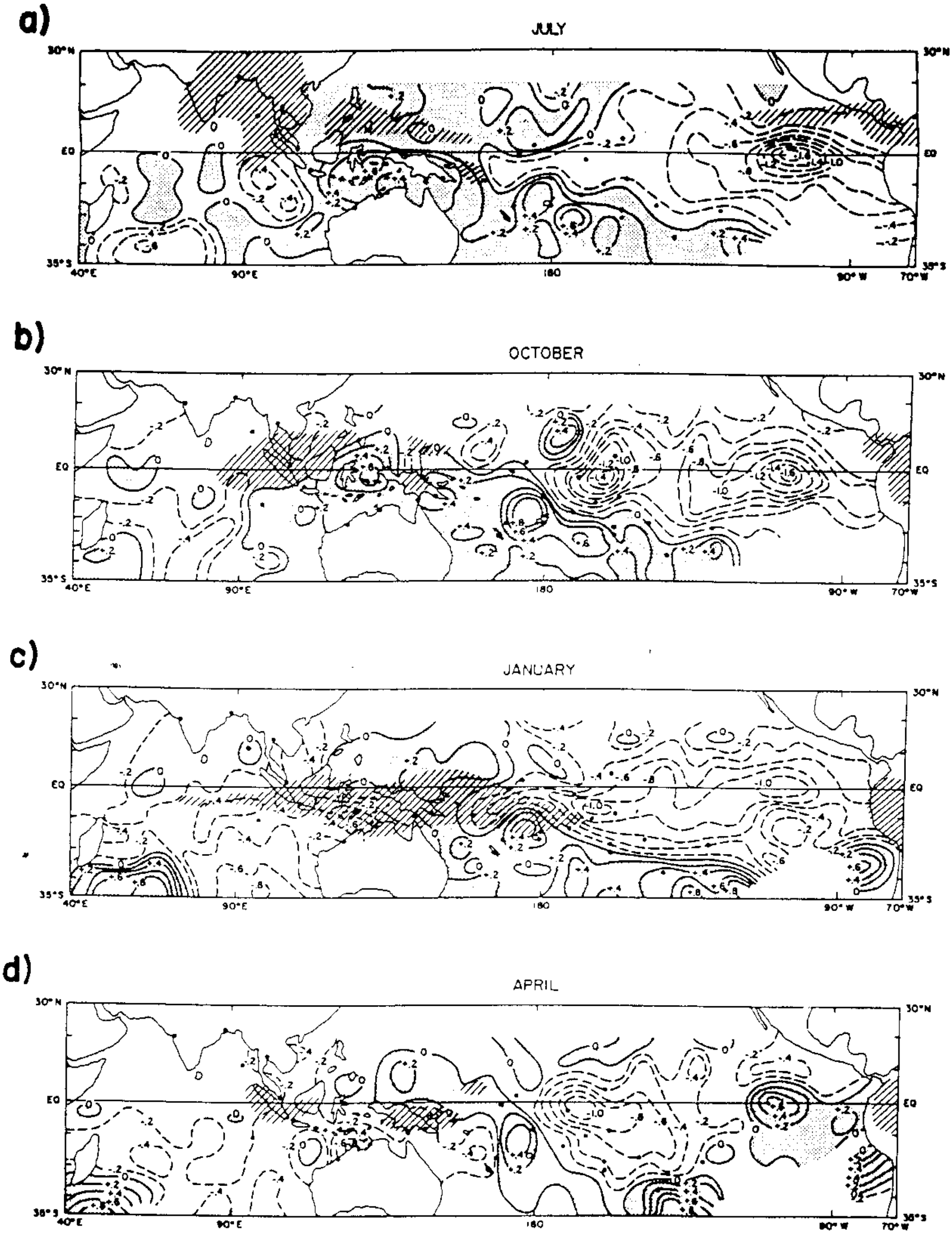
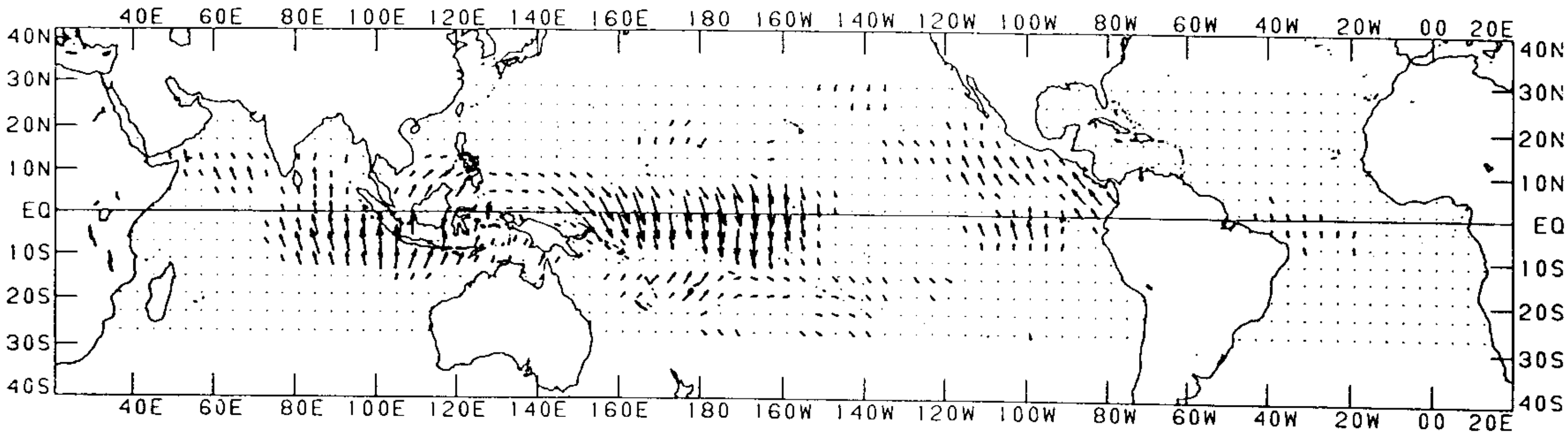


~~Fig. 6~~. Long-term mean OLR values less than 220 W m^{-2} indicative of areas of heaviest convection (hatching) and SST differences, strong years minus previous years for (a) July, (b) October, (c) January, and (d) April (SST values after Meehl 1987). OLR values from Janowiak et al. (1985).

Fig. 1

Fig. 1



~~Fig. 13~~. Biennial coherence between the sea surface temperature in the central Pacific (equator, 171°W) and the zonal wind component over the global oceans represented by harmonic dial vectors. The maximum biennial SST–zonal wind correlation (0.78) is at the base point. The downward (upward) pointing arrows signify an in-phase (one-year out-of-phase) relationship.

Fig. 2

Fig 3.

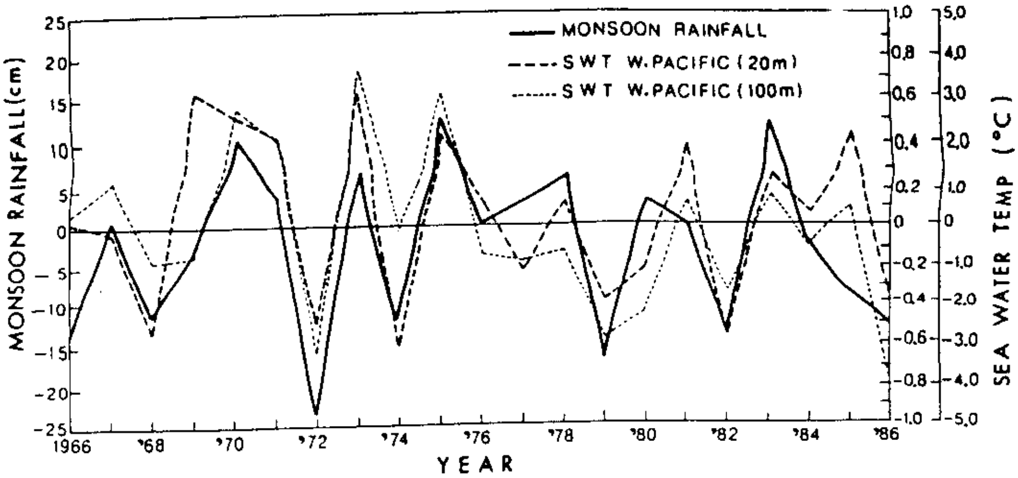


FIG. 3. Time series of Indian monsoon rainfall anomaly (thick solid line) and SWT anomaly at 20 m (thick dashed line) and 100 m (thin dashed line) depth averaged for the 137°E line (2°N–10°N) in the succeeding January (Yasunari 1990).

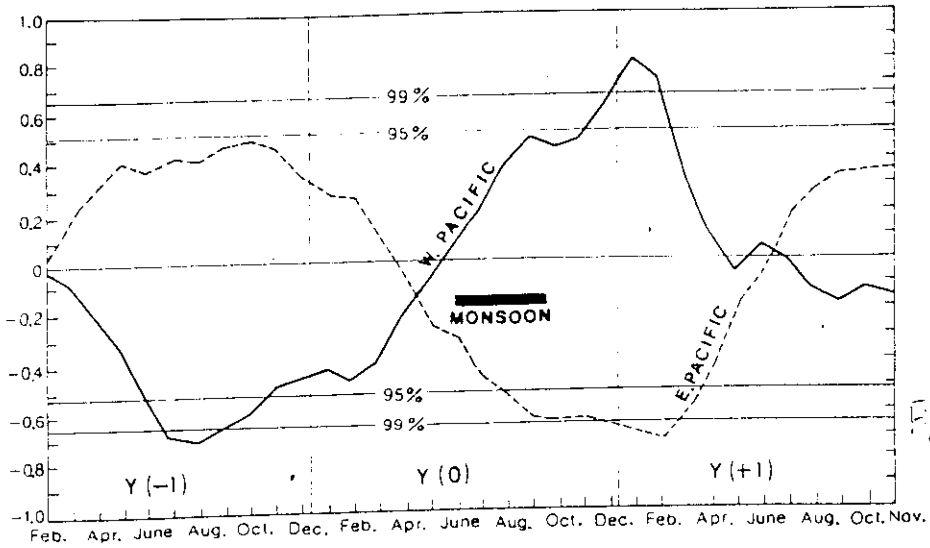


Fig 4

FIG. 4. Lag correlations between Indian monsoon rainfall anomaly and SST anomaly in the western (0°–8°N, 130°E–150°E) and the eastern (0°–8°N, 170°W–150°W) Pacific. The reference monsoon season is shown with thick black bar. Y(0) denotes the year of reference monsoon and Y(-1) [Y(+1)] denotes the year before (after) Y(0) (Yasunari 1990).

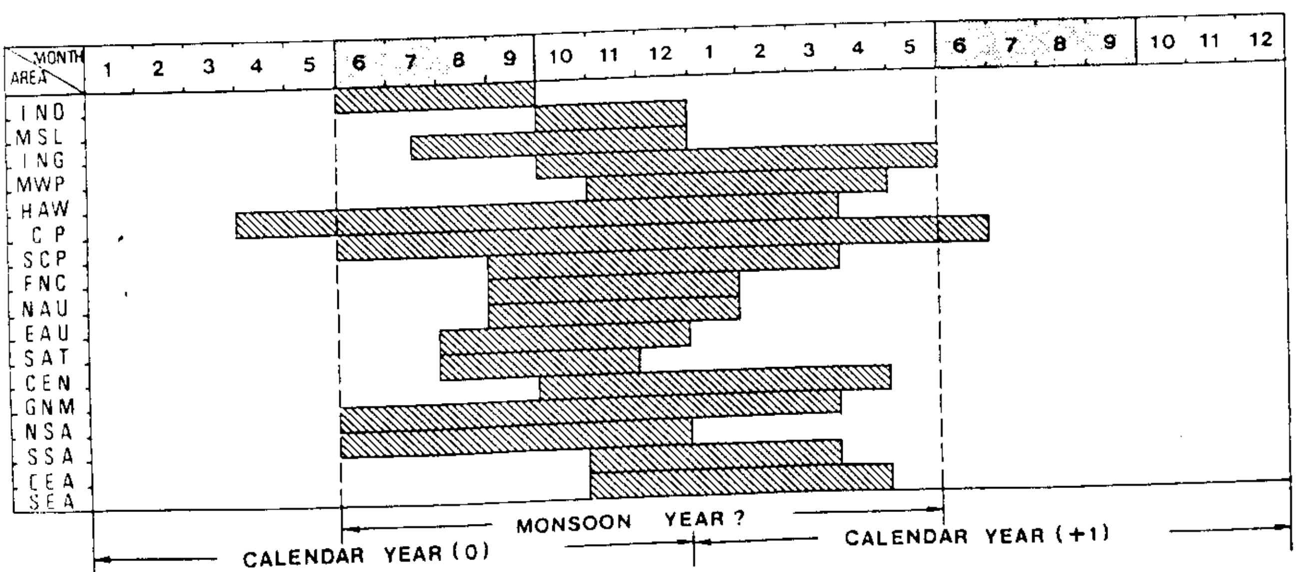


Fig 5

Fig 5

~~Fig. 6~~. **Seasons of precipitation anomaly related to the extreme phases of SOI for the areas in the tropics and subtropics. Dashed vertical lines indicate the approximate extent of the "monsoon year."** Abbreviated area names are as follows: IND, India; MSL, Minicoy/Sri Lanka; ING, Indonesia/New Guinea; MWP, Micronesia/West Pacific; HAW, Hawaiian; CP, Central Pacific; SCP, South Central Pacific; FNC, Fiji/New Caledonia; NAU, Northern Australia; EAU, Eastern Australia; SAT, Southern Australia/Tasmania; CEN, Central America/Caribbean; GNM, Gulf/North Mexico; NSA, Northeastern S. America; SSA, Southeastern S. America; EEA, Eastern Equatorial Africa; SEA,

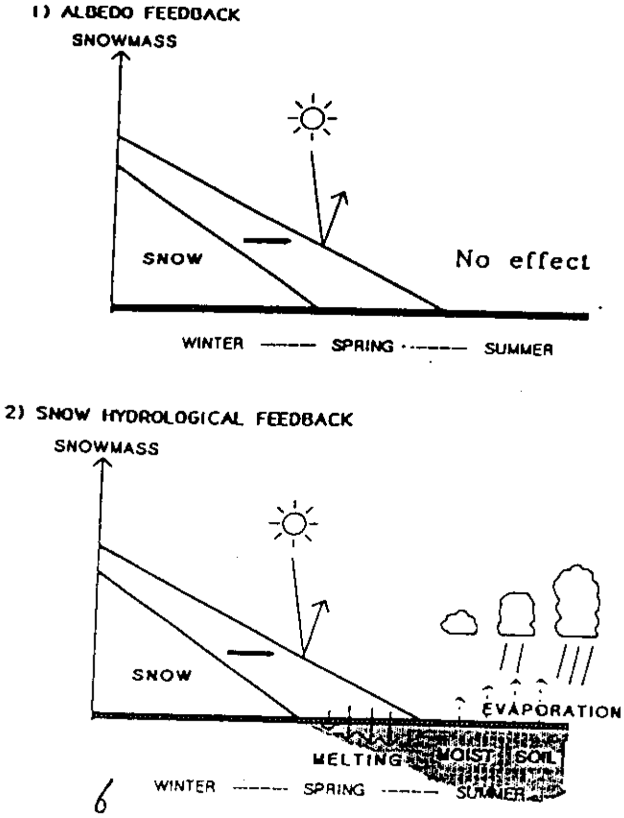


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Fig. ~~4~~. Schematic diagram for the albedo feedback and the hydrological feedback of snow cover during the seasonal march from winter to summer.

Fig 6

Fig 7

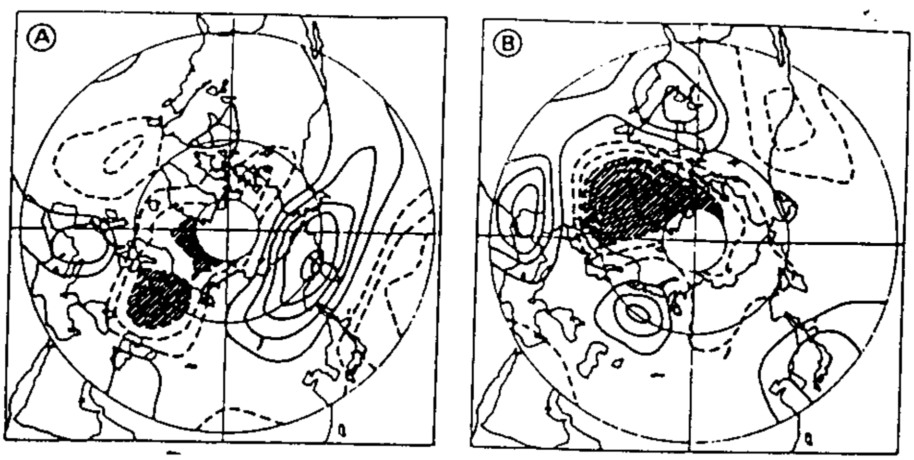
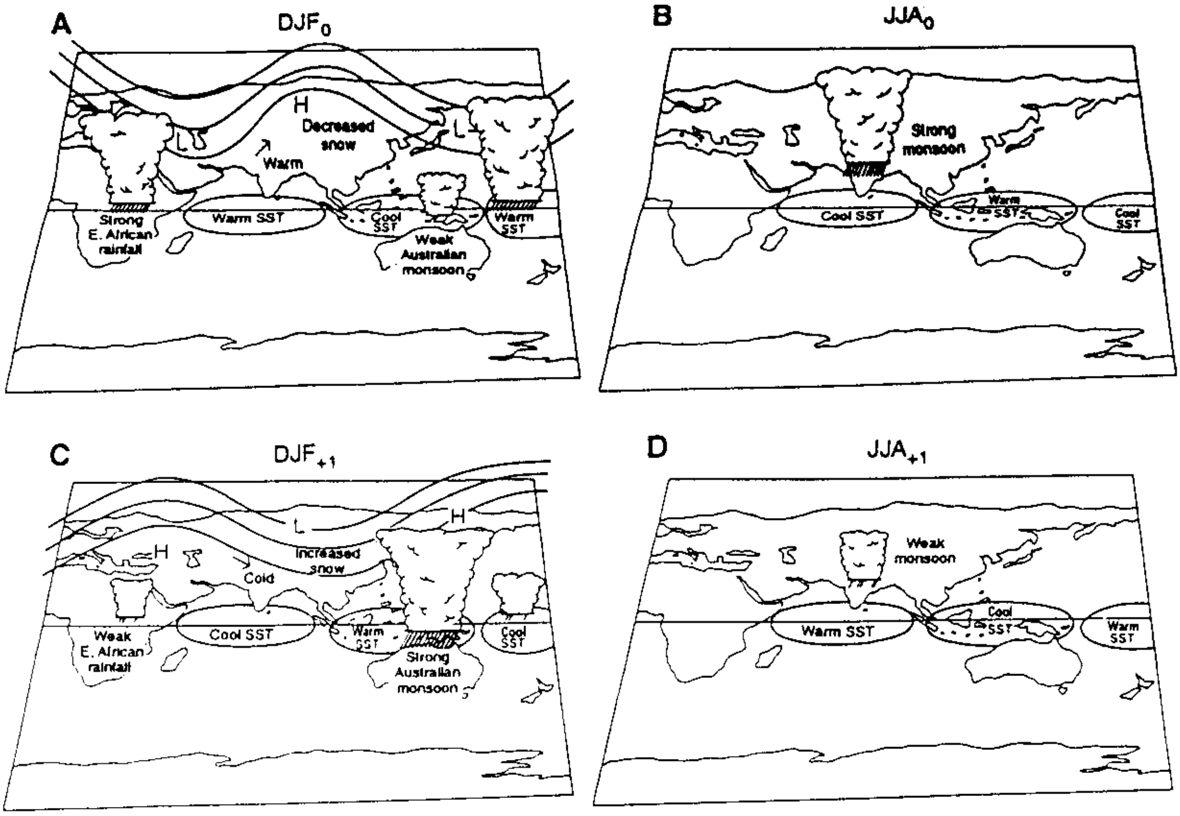


Fig. 11. Composite anomalies of 500 mb geopotential height ~~from August to November~~ for December of strong (left) and weak (right) Indian monsoon years. Contours are 10 gpm and negative values are shown with dashed lines.



~~Fig.~~ 3. Schematic biennial evolution from the northern winter before a strong Asian monsoon (**A**) through the strong monsoon season (**B**), to the northern winter after the strong monsoon before a weak monsoon (**C**), to the following weak monsoon (**D**).

Fig. 8

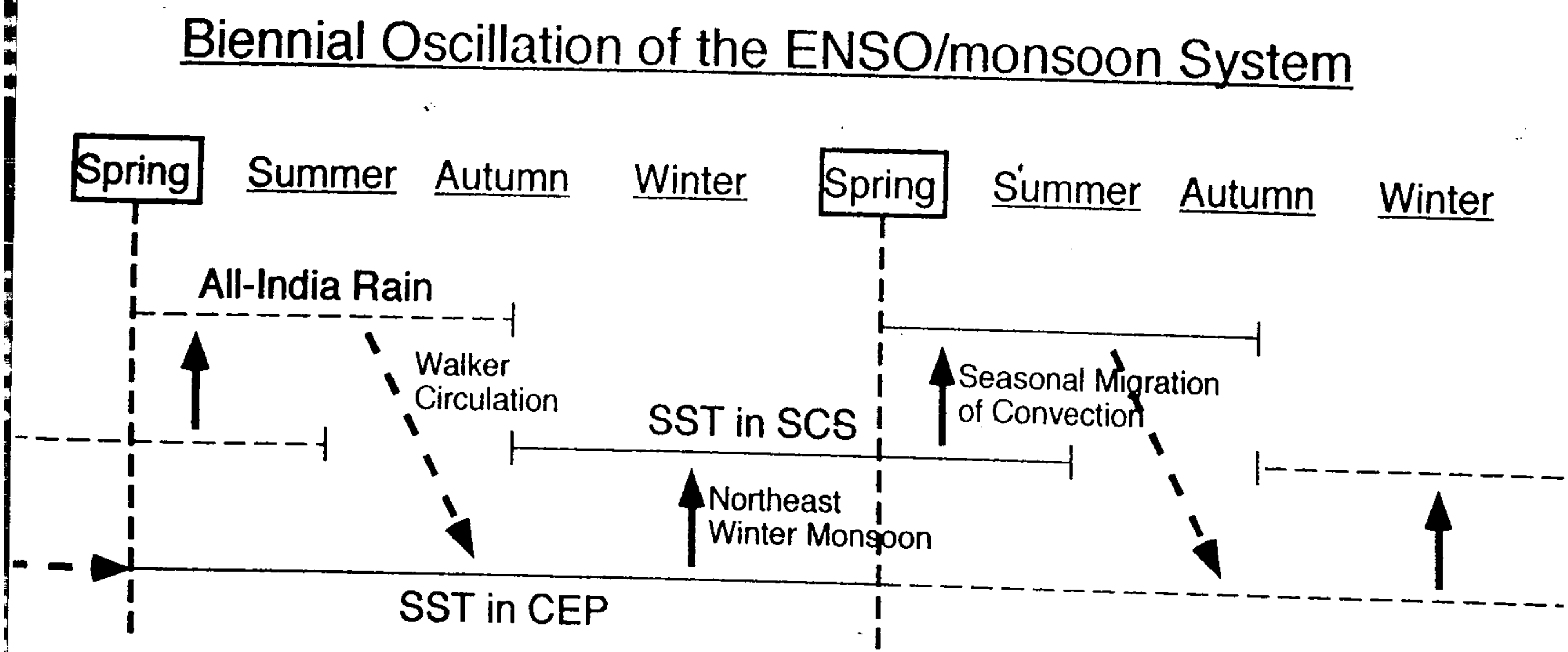


Fig. 9

A schematic diagram illustrating the biennial cycle of the ENSO/monsoon system. The thin solid and dashed lines indicate the periods of positive and negative anomaly, while the thick solid and dashed vectors indicate the positive and negative correlations, respectively. Abbreviations SCS and CEP denote the South China Sea and the central equatorial Pacific.