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Anthropogenic Climate Change Prediction, Detection and Attribution

The work of the Intergovernmental Panel on Climate Change is underpinned by CLIVAR's research on Detection and Attribution of Climate Change. This special issue focuses on these topics.

Higher Risk of wet Winters in Europe due to Climate Change ?

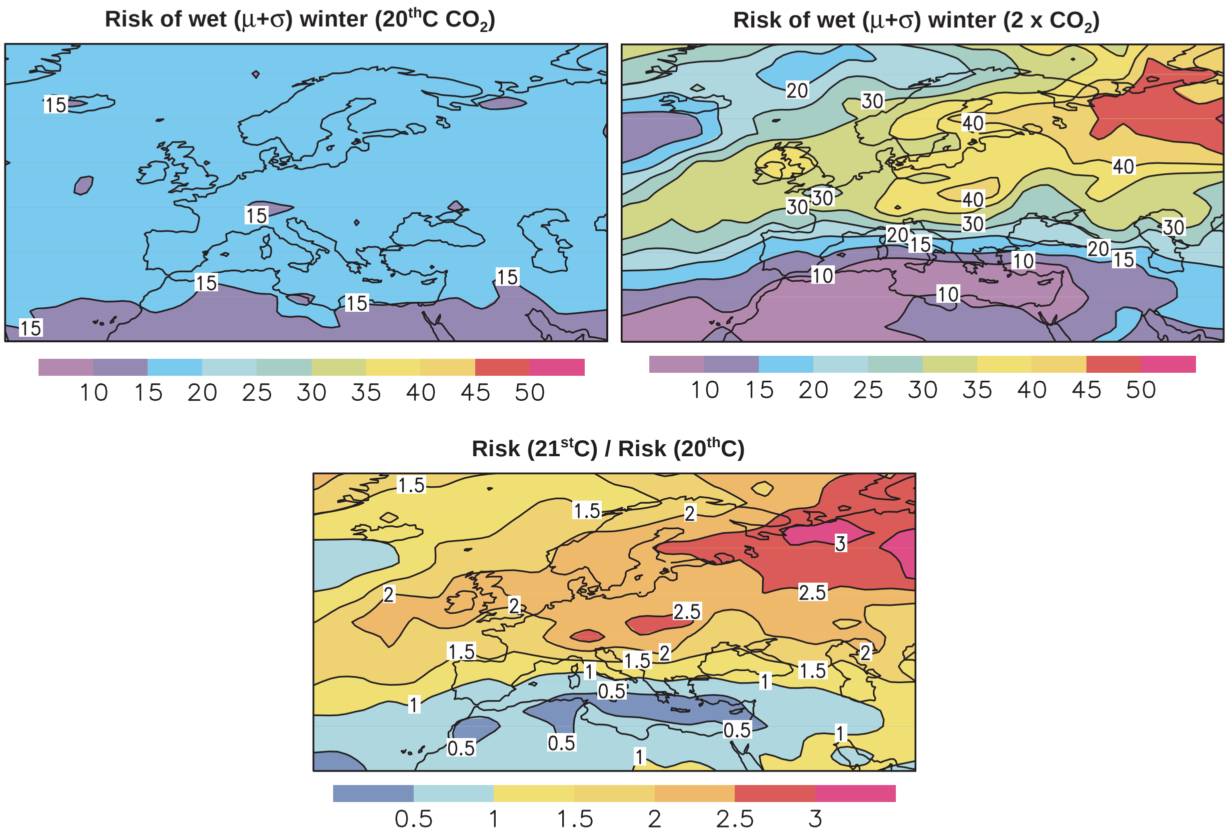

Figure 1 from paper 'Quantifying risk in a changing climate': by T. Palmer and J. Räisänen:
Upper left: the probability of a wet winter defined from the control CMIP2 ensemble with 20th century levels of CO_2 and based on the event E: DJF rainfall greater than the mean plus one standard deviation. Upper right: the probability of E but using data from the ensemble with transient increase in CO_2, and calculated around the time of CO_2 doubling (years 61-80 from present). Bottom: the ratio of the middle to top panel values, giving the change in the risk of a wet winter, arising from man's impact on climate. The paper appears on page 3.

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Editorial

This year, the Third Assessment Report (TAR) "Climate Change 2001" of the Intergovernmental Panel on Climate Change (IPCC) (http://www.ipcc.ch) has been published. Its three volumes deal with the "Scientific Basis", "Impacts, Adaptation and Vulnerability" and "Mitigation". (These are supplemented by more widely distributed "Summaries for Policymakers"). The TAR represents an enormous effort by many scientists in collating, reviewing and presenting the present state of our knowledge of the climate system, of the known and potential human influences and of the likely global and regional impacts of climate change.

When CLIVAR was originally conceived it was structured (for the sake of convenience) around three timescales - Seasonal to Interannual (dealing with phenomena such as ENSO and Monsoons), Decadal to Centennial and Anthropogenic Climate Change. The work of IPCC is clearly focused on Anthropogenic Change issues but, because of the short (typically only 100 years) instrumental record available to us, understanding the inherent variability of the climate system even on interannual scales (Do ENSO events introduce regional and global biasses?) is crucial. Thus all of CLIVAR's observational and modelling activities underpin the IPCC process.

We have therefore taken this opportunity to include in this issue of CLIVAR Exchanges some articles that are closely related to the IPCC activities. We had a very good response for our call for papers and include articles on various facets of the problem ranging from the investigation of anthropogenic, influences on natural climate phenomena to the detection of anthropogenic changes in the ocean and the means by which the probability of extreme events might be quantified (cover pictures).

As more and more components of CLIVAR progress from planning to implementation so the organizational structure is growing. A Southern Ocean Implementation panel has been added and a Pacific Panel will shortly join it. The Asian-Australian Monsoon Panel just held its first meeting under its new co-chairs and this meeting is notable in including participation from representatives of the VAMOS (American monsoon system), African Climate Variability panels and of the WCRP's GEWEX (Global Energy and Water Experiment) project. All these elements are to a great extent interdependent.

In order to inject fresh and innovative ideas into CLIVAR oversight and to involve as many scientists as possible from many nations and institutions involved in CLIVAR science, CLIVAR panels rotate their membership. Thus, we welcome a number of new members in the CLIVAR Scientific Steering Group: Dr. Ian Simmonds, School of Earth Science, U. Melbourne, Australia, Dr. Pedro Silva Dias, Instituto Astronómico e Geofísico, Universidade de São Paulo, Brazil, Dr. Max Suarez, NASA/GSFC, Greenbelt, USA and Dr. Kensuke Takeuchi, Hokkaido University, Sapporo, Japan. Short introductions with a CV and their main research interest and expertise from Ian and Kensuke can be found on page 25. While welcoming these new members we express our thanks to the outgoing members Dr. Edward Sarachik (a founder member of the SSG), Dr. Neville Nicholls (who continues as a member of the A-A Monsoon Panel , Dr. John Mitchell (a key player in the IPCC process) and Dr. Kimio Hanawa, Tohoku University, Japan for their work as members of the CLIVAR SSG.

Thanks to funding from the US CLIVAR Interagency Group (IAG) we have also been able to hire a second new staff member in the ICPO. Dr. Daniela Turk will join the office in mid-September. Daniela is currently at Dalhousie University in Halifax, Canada where she has been studying the biogeochemical impacts of ENSO in the Pacific with Drs Marlon Lewis and Mike McPhaden. She will assume responsibility in the ICPO for Pacific Issues and for CLIVAR's links with the International Geosphere-Biosphere Programme. Also assisting CLIVAR implementation is Dr Mike Sparrow. He has joined the WOCE IPO with primary responsibility for the production of a series of Atlases of WOCE data. Mike spent the past 5 years on the analysis and interpretation of subsurface lagrangian data from the North Atlantic and in light of his earlier research on Southern Ocean circulation will support the Southern Ocean panel.

The co-ordinated international projects planned under CLIVAR represent the integration of many national research projects. Information about funded and implemented projects is available on the WWW through CLIVAR's Searchable Programme Information Network (SPRINT) database. We encourage you to explore this by logging on to http://clivar-search.cms.udel.edu/projects/index.htm The database is only as good as the information we receive and so Katherine Bouton in the ICPO is looking forward to your comments, additions or corrections.

Programming has just started for CLIVAR's first Science Conference in late 2003 / early 2004. This will provide the opportunity for an overview of all the science falling within CLIVAR's wide remit. The Scientific Steering Group are pleased that Prof. Lennart Bengtsson, former director of the department for climate modelling at the Max-Planck-Institut für Meteorologie in Hamburg and one of those who had an early influence on CLIVAR's direction, has agreed to chair the scientific organizing committee.

We hope that you will enjoy this issue of Exchanges. The next one to be published at the end of the year will promote the newly developing Southern Ocean focus of CLIVAR.

John Gould and Andreas Villwock

Quantifying risk in a changing climate

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The autumn and winter of 2000/2001 were the wettest on record over England and Wales, with widespread flooding. Throughout the period, there was a general sense of concern amongst society at large, that such heavy and prolonged rainfall was attributable in some measure to anthropogenic global warming. Indeed government ministers referred to these events as providing a 'wake-up call' to man's effect on climate. On the other hand, a commonly-held view amongst climate scientists is that (notwithstanding the profound effect that global warming is likely to have on climate) it is impossible to attribute individual climate anomalies to the influence of anthropogenic forcing.

However, it is argued here that probabilistic risk analysis applied to muti- climate-model ensemble projections, provides a scientifically-sound methodology to quantify the role of anthropogenic forcing on specific climate anomalies (Räisänen and Palmer, 2001), in such a way as to address these societal concerns. Such analysis is becoming common place in weather and seasonal climate prediction (Palmer, 2000), though has not yet formed part of routine climate-change assessments e.g. as presented in IPCC (2001). Of course the analysis here is not specific to UK flooding, and provides a general methodology for quantifying changes in climate risk.

The basis of the method relies on being able to quantify the essential characteristics of the specific climate anomaly of interest in terms of a relatively rare climate event *E*. Here *E* is based on seasonal statistics, so that, over any particular season, *E* either occurs, or does not occur. For example, the 2000/2001 flooding over the UK was associated with a very substantial projection of the atmospheric circulation onto a naturally-occurring pattern of interannual variability (the so called Eurasia-1, or Scandinavian teleconnection pattern (Barnston and Livezey, 1987; Blackburn and Hoskins, 2001; see also see http://www.cpc.noaa.gov/products/), with streamfunction anomalies (of alternating sign) over the eastern Atlantic, United Kingdom and northern Scandinavia. (The fact that circulation anomalies project substantially onto naturally-occurring patterns of variability is not an argument that such anomalies cannot be attributable to anthropogenic forcing; Palmer, 1999; Corti et al., 1999). As such, from a purely meteorological point of view, *E* could be defined by the condition that the seasonal-mean projection coefficient of the Eurasia-1 anomaly pattern exceeded (say) 2 standard deviations. From a more practical perspective *E* could be defined by the physical condition of specific rivers bursting their banks during the season of interest.

Let *f1(E)* and *f2(E)* denote the frequency of occurrence of E in a climate with single and doubled values of 20th Century CO_2 content. One could perhaps imagine *f1(E)* to be defined by 20th Century climate statistics, whilst *f2(E)* is estimated from an integration of a comprehensive climate model with projected levels of CO_2 (and coupled to a hydrological model if *E* is explicitly defined by fooding events). In practice, estimates of the frequency of relatively rare events cannot be reliably sampled from short timeseries, irrespective of whether such timeseries come from observations or model integrations. Secondly, estimates of the projected probabilities will be sensitive to uncertainties in the computational representation of unresolved scales in the climate equations.

In practice the effects of both sampling and model uncertainty can be taking explicitly into account by estimating both *f1* and *f2* from multi-model ensembles - here they are based on estimates from multi-model ensemble integrations from CMIP2 (Second Coupled Model Intercomparison Project), comprising integrations from 13 state of the art coupled ocean-atmosphere climate models (Meehl, 2000). The first (control) ensemble was run with 20th Century CO_2; the second with a transient compound increase in CO_2 of 1%/year. In view of the limited ensemble size available, and since hydrological models (with detail pertaining to specific river basins) have yet to be driven with output from this type of ensemble, in practice *E* is based neither on the Eurasian-1 pattern, nor on river flooding, but on seasonal-mean precipitation. Specifically, *E* is said to have occurred if December-February mean precipitation exceeds one standard deviation (σ) above normal (μ). For estimating both *f1* and *f2*, σ and μ are defined (at each grid point) by the CMIP2 ensemble of control integrations.

Fig 1a (page 1) shows *f1* over Europe from the control ensemble. As would be expected, there is an approximately 16% probability of occurrence of *E*s. Fig 1b shows the equivalent estimate of *f2* from the climate change ensemble taken from years 60-80, around the time of CO_2 doubling. It can be seen that over much of central and northern Europe, the probability of *E* has increased substantially, consistent with a likelihood of enhanced storm-track activity. By contrast, over parts of the Mediterranean and Northern Africa, the probability of *E* has decreased.

Fig1c shows the ratio *f2/f1* which can be interpreted as giving a measure of the effect of anthropogenic forcing on the risk of *E*. Over parts of northern Europe (including much of the United Kingdom) the risk of *E* has approximately doubled at the time of CO_2 doubling. By contrast over parts of the Mediterranean the chance of *E* has approximately halved. Although *E* is not sufficiently extreme

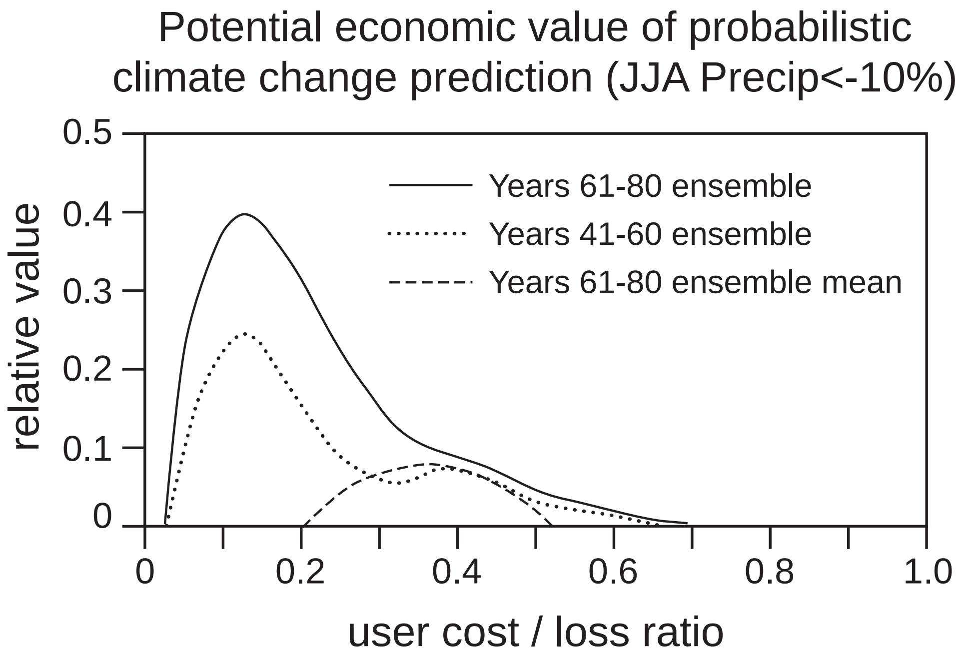

Fig. 2: Potential economic value of the CMIP2 multi model climate change ensemble evaluated by choosing, at random, one member of the ensemble as 'truth'. Based on the event: June-August precipitation less than -10% of the control climate ('dry summer'). Dotted: value at years 41- 60 of the transient CO_2 run. Solid: years 61-80 of the transient CO_2 run. Dashed: ensemble mean of years 61-80.

that it can be said to describe flooding, this analysis permits the newsworthy yet scientifically-justifiable assessment: '*The risk of wet winters over the UK will double over the next half century, due to man's effect on climate*'.

The representation of climate change in terms of (a probabilistic) risk assessment differs from the more conventional approach of collapsing information from an ensemble into a deterministic ensemble-mean prediction (e.g. IPCC, 2001). It is contended that such deterministic predictions have less potential economic value that the type of probabilistic assessment presented here. To make this claim, a decision-analytic model used to assess the utility of ensemble weather forecasts and seasonal predictions (Richardson, 2000; Palmer et al., 2000) has been adapted. An example of the application of this decision analytic model (from Räisänen and Palmer, 2001) is shown in Fig 2. A member is randomly chosen from the ensemble, and is assumed to represent 'truth'. The normalised economic value of decisions made using probabilistic projections from the full CMIP2 ensemble can be assessed as a function of the so-called cost-loss ratio of the user. Fig 2 shows that this value exceeds that based on decisions made using the deterministic ensemble- mean projection. The reason is that ensemble-mean forecasts tend to hedge towards the climate mean state, and underpredict the frequency of extreme events. We refer to Räisänen and Palmer, (2001) for more details.

In conclusion, it is recommended that climate change projections be based on probabilistic risk assessment, and move away from the more traditional deterministic ensemble mean approach.

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Predictability of the Atlantic Thermohaline Circulation under Global Warming

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1. The fate of the THC under global warming

In the recently published Third Assessment Report of the Intergovernmental Panel of Climate Change (IPCC, 2001) it was concluded that '*most models show a weakening of the ocean thermohaline circulation*' and that '*beyond 2100, the thermohaline circulation could completely, and possibly irreversibly shut down ...*'. The possiblity of such reorganisations represents additional and considerable uncertainty in climate projections of the future. CLIVAR is expected to contribute to a better understanding of these issues both through CLIVAR's 'Anthropogenic Climate Change' and its research on 'Atlantic Thermohaline Circulation'.

High-resolution paleoclimatic records (Broecker, 1997; Stocker, 2000), as well as numerous model simulations (an overview is given by Cubasch et al., 2001) suggest that global warming may reduce the merdional overturning circulation in the Atlantic (thermohaline circulation, THC). Some climate models even show a complete shut-down of the THC if some critical thresholds are crossed (Manabe and Stouffer, 1993; Stocker and Schmittner, 1997). However, the range of possible responses for the next 100 years is rather large: the THC in some models remains almost constant (Latif et al., 2000; Gent, 2001), while most models exhibit a substantial reduction of the meridional overturning. This uncertainty is likely due to model differences in climate sensitivity, in the response of the hydrological cycle, and in the representation of processes and feedbacks (e.g., NAO, ENSO). This indicates that the strength of stabilising and destabilising feedbacks influencing the THC is still largely unknown. Current research, therefore, must be concerned with a better understanding of the underlying processes.

However, there is an additional uncertainty which may severely limit the long-term predictability of the Atlantic THC. This uncertainty is associated with the fact that thresholds of instability in the THC may exist and that close to such thresholds prediction may be greatly reduced. The following is a brief summary of a paper in press (Knutti and Stocker, 2001).

2. Ensemble simulations below the instability threshold

We use a highly simplified coupled climate model which consists of a zonally averaged ocean model coupled to a moist energy balance model of the atmosphere, and includes the seasonal cycle (Schmittner and Stocker, 2001). Due to the absence of atmospheric dynamics and its associated variability, we employ a weak white-noise freshwater flux perturbation at the ocean surface to mimick the effect of 'weather' on the THC. Because the coupled model is efficient, Monte Carlo simulations can be performed. A ensemble consists of 100 simulations which differ only by the random sequence of freshwater flux perturbations. We assume a simple global warming scenario, in which the CO_2 is doubled within 140 years (0.5% increase per year) and held constant thereafter. Climate sensitivity is a parameter in this simplified climate model which is selected. Although internal variability is present in the model, the global mean temperature is tightly constrained (Fig. 1a). Global warming leads to a reduction in the strength of the Atlantic THC, and a new steady state is approached (Fig. 1b). The reduction is due to the combined effect of warmer

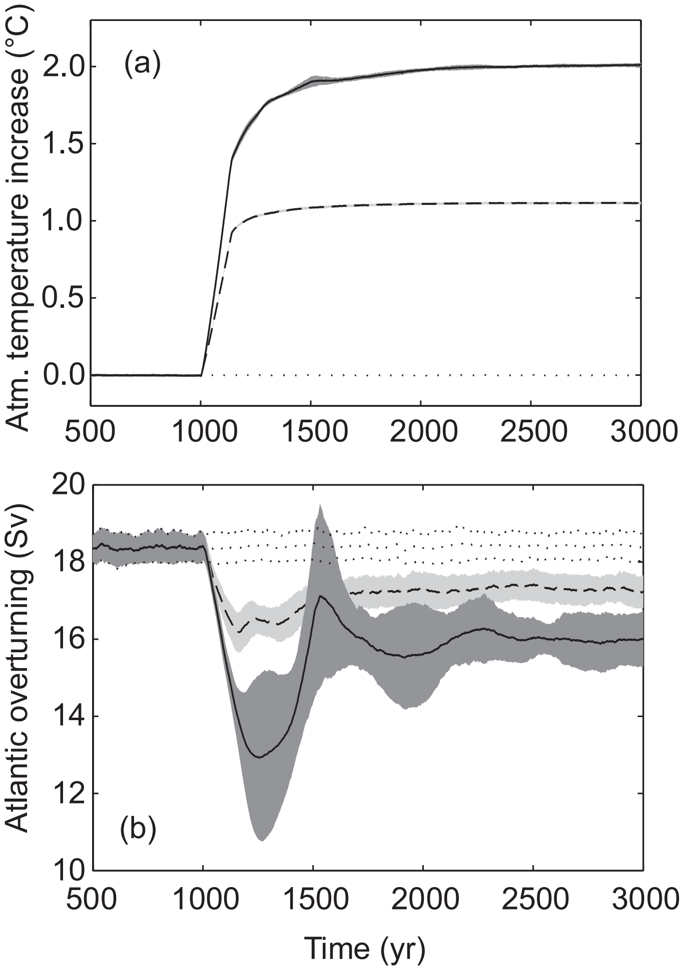

Figure 1: (a) Globally averaged surface temperature simulated by the coupled model of reduced complexity for two cases of prescribed climate sensitivity (1°C and 2°C for a doubling of CO_2 within 140 years). (b) Evolution of the overturning in the Atlantic (in Sv, 1 Sv = $10^6 m^3 s^{-1}$) for the weak (dashed) and the intermediate (solid) climate sensitivity, and a control simulation (dotted). The shaded bands denote the 1-σ range determined from 100 ensemble simulations with random sequences of freshwater anomalies (from Knutti and Stocker, 2001).

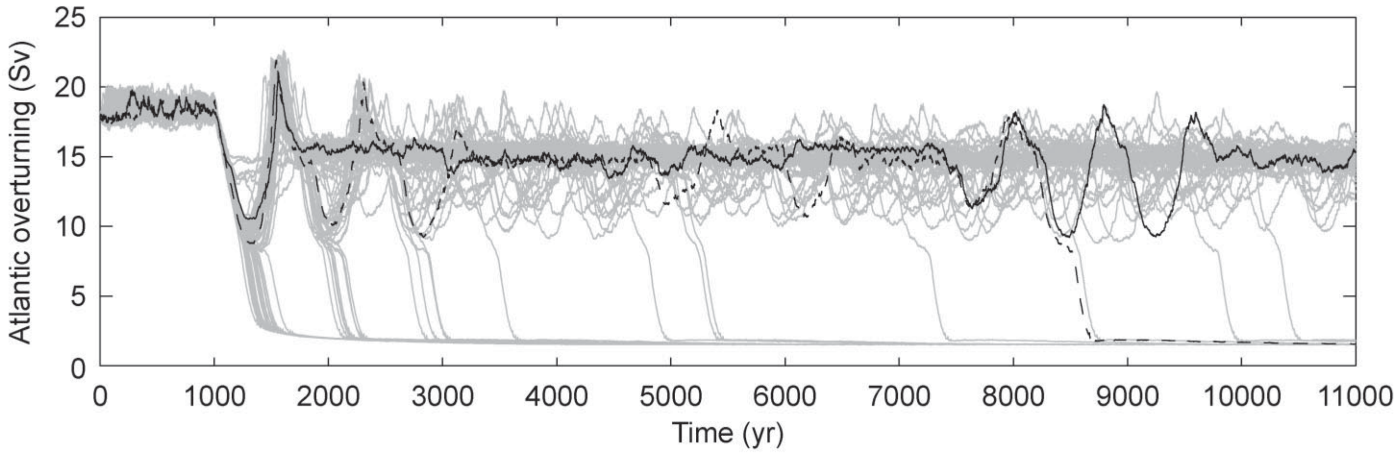

Figure 2: Multi-millennial evolution of the Atlantic overturning for 100 simulations (gray lines) using stochastic forcing close to an instability threshold of the thermohaline circulation. Complete shut-downs of the thermohaline circulation may occur even long after the radiative forcing has stabilised to a new mean value (by year 1140). Two simulations with qualitatively different behaviour are highlighted as black lines (from Knutti and Stocker, 2001).

sea surface temperatures and a stronger hydrological cycle consistent with comprehensive 3-dimensional climate models (Dixon et al., 1999; Mikolajewicz and Voss, 2000).

The present model posesses a second equilibrium circulation in which the Atlantic THC is shut down, i.e. there exists an instability threshold. When the Atlantic overturning reduces, the circulation approaches this instability threshold. The character of the transient phase is strongly dependent on how close the circulation comes towards the threshold (Fig. 1b). For the smaller climate sensitivity, the range of different evolutions within the ensemble (shaded band) remains approximately constant. Closer to the instability, the range grows significantly, even though the amplitude of the random forcing is unchanged. Even long after the perturbation, the ensemble range remains increased. It is also evident from Fig. 1b, that the predictability of the THC is particularly reduced during the transient phase.

3. Crossing the instability threshold

By increasing the climate sensitivity further, some of the ensemble members may now exhibit a full shut-down of the THC (Fig. 2). Most of the transitions occur at the time when the THC strength is most reduced, consistent with indications from 3-dimensional models (Tziperman, 1997). In realisations where the THC recovers, there is a tendency to multi-century oscillations of the Atlantic overturning. These oscillations seem to be damped on the time scale of a few thousand years, but there are also cases, where these oscillations re-appear much later (solid line) after a quiet period of many millennia. When the system is close to the instability, transitions can occur long after the perturbation. This implies a complete loss of predictability for the THC strength close to the threshold.

4. Conclusions

Investigating the long-term fate of the THC requires multi-century integrations of climate models. The currently available computing capacity still excludes thorough parameter studies with comprehensive 3-dimensional climate models. An assessment of uncertainty is therefore impossible. Models of reduced complexity can help overcome this difficulty and make valuable contributions to a better understanding of parameter space. These models are most useful as exploratory tools for hypothesis building. New concepts, or new research strategies can be readily assessed with these models. However, such models, which are intermediate within the climate model hierarchy, must be used wisely if they are to contriubute to scientific progress. In a continuous mode of scientific exchange, results from these models must be critically compared with those from comprehensive climate models in order to determine the robust findings.

It is herefore of particular importance in hierarchical climate modelling to test the new insights with specifically designed experiments using comprehensive models. Hopefully, multi-century ensemble simulations investigating the fate of the THC will soon become feasible.

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Can coupled models help to define an observing strategy for detection of climate change in the ocean?

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One of the priorities of the Global Climate Observing System is to make observations which allow the earliest possible detection of climate trends and climate change due to human activity. Until now detection of anthropogenic climate change has focused almost exclusively on surface temperature (for example, Barnett et al., 1999). In order to increase our confidence in these results it is desirable to apply this technique to other aspects of the climate system. Detection and attribution is also important in terms of model validation as it allows us to test the ability of models to reproduce historical changes. In the case of tropospheric temperature changes, discrepancies between observed and modelled changes have not yet been fully explained (IPCC, 2001).

We might expect the ocean to be less noisy than other parts of the climate system and for changes in surface fluxes (dynamic and thermodynamic) to change the ocean properties producing high signal-to-noise ratios. While decisions on where to make observations can often be made on a subjective basis we here attempt to make an objective assessment of which ocean variables will be useful for detection of anthropogenic climate change. Santer et al. (1995) looked at the signal-to-noise ratios of a number of different ocean indices in a coupled model and ocean-only models. We have extended this idea using HadCM3, one of the first coupled climate models to be run without artificial flux adjustments (Gordon et al., 2000). Clearly there are other requirements for ocean observations: forecasting, abrupt climate change, etc but our aim here is to show the potential usefulness of an objective assessment.

We make use of two experiments performed with HadCM3: CTL, the control experiment with fixed preindustrial greenhouse gases, and B2, an anthropogenic experiment forced with historical greenhouse gases, etc until present day and projected to 2100 with the IPCC SRES B2 scenario (IPCC, 2001). For a range of indices we have assumed that the index would be measured once and then for a second time fifty years later. We have calculated all the fifty year (overlapping) changes in each index from both CTL and B2 and compared the two distributions. It is clear when an index has a high signal-to-noise ratio because there is a clear separation of the two distributions.

Figure 1 shows the distributions from four indices; global mean SST, heat transport and the overturning streamfunction at 1500 m at 24°N in the Atlantic, and SubAntarctic Mode Water (SAMW) salinity. The distributions show that measures of the temperature and salinity of the ocean have high (and significant) signal-to-noise ratios on these timescales, the ratios of measures of the circulation (streamfunction) are marginally significant while for heat transports the two distributions are not distinguishable. This suggests that observations of heat transport are unlikely to be useful for the detection of climate change and we would need to measure the circulation over long periods. Ocean temperature and salinity on the other hand may have the potential for detection of change.

To examine this more closely, we have calculated signal-to-noise ratios in the temperature and salinity of each ocean basin. If we look on pressure surfaces we generally see fairly high signal-to-noise ratios in temperature but not salinity. To combine the two measurements we can look at salinity (or temperature) changes on isopycnal surfaces. Figure 2 shows where the signal-to-noise ratios are significant for the average fifty year signal from B2. These results show that there are large regions of the ocean where sig-

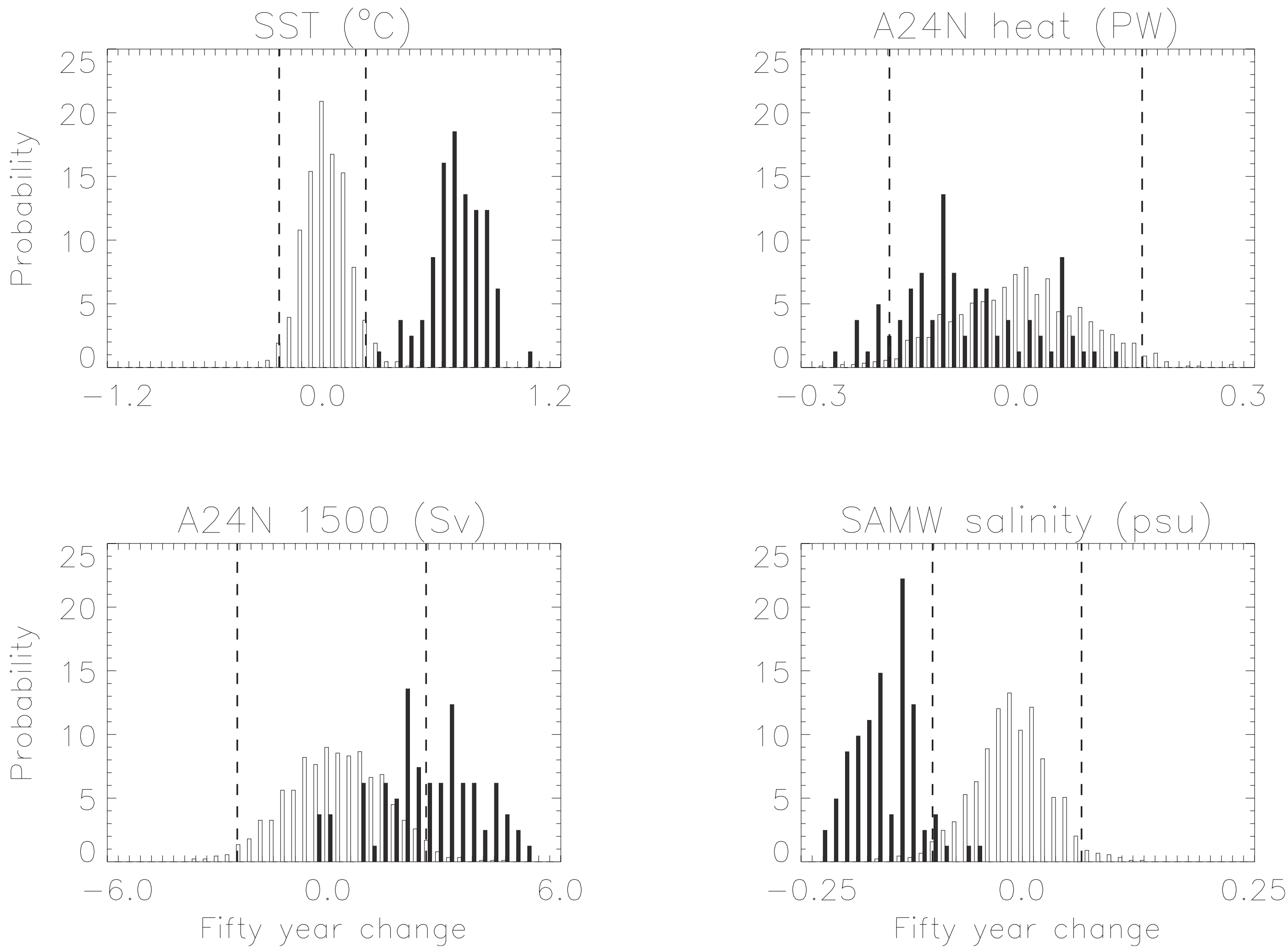

Figure 1: Distributions of fifty year changes in each variable. The unfilled distribution is from 940 years of CTL, the filled distribution from B2 years 1970-2100. The distributions are plotted in terms of their probability. The dashed lines indicate the 5% significance levels assuming a two-tailed test.

nal-to-noise ratios of temperature-salinity properties may be high (e.g., in SAMW as shown by Banks et al., 2000).

In summary, our results show that we can objectively analyse output from a coupled climate model to provide guidance on which ocean observations will be most useful in terms of detecting anthropogenic climate change. Temperature and salinity properties appear to provide the most realistic means of detecting change in the ocean. Our early results suggest that the Argo programme combined with historical hydrographic data (for example, Levitus et al., 1998) and repeat sections along some of the WOCE lines would provide us with the measurements which are needed to detect climate change. To establish confidence in results such as these we need to understand model processes (e.g., Banks et al., 2001) and validate the internal variability of the model.

Although this analysis suggests that measurements of the thermohaline circulation will detect climate change at a lower significance level, it is possible that these measurements will be important for monitoring of abrupt climate change. In the future we plan to develop this approach to provide guidance on the suitability of observations for other requirements.

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Sensitivity of permafrost cover in the Northern Hemisphere to climate change

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In the Northern Hemisphere (NH) permafrost covers about 25.5 million km^2 including 11.7 million km^2 covered by continuous permafrost (Anisimov and Nelson, 1997), 25% and 12% of the total NH land area. Being located in the northern parts of NH continents, permafrost cover can be rather sensitive to anthropogenic climate changes. Permafrost cover study is one of the priorities for the Russian CLIVAR activity (Gulev and Mokhov, 1999; Mokhov, 1999).

To compare modelled and reconstructed permafrost changes, at a first step it is helpful to use the simplest climatic characteristics. These characteristics should be independent of details of soil process parameterizations and ideally be tested against both instrumental and paleoclimatic data. Some of such indices are based on surface air temperature. Actually, they characterize climatic conditions favourable for permafrost formation.

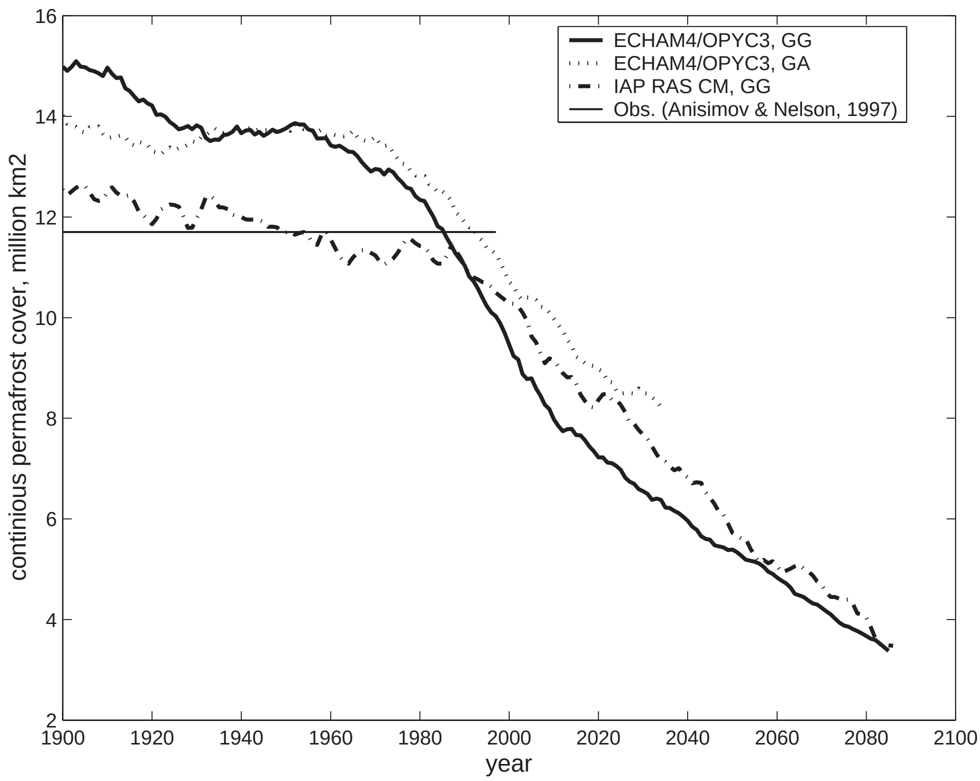

Fig.1: Diagnosed continuous permafrost cover, million km², 30-year moving average.

Here NH PC sensitivity is estimated using the numerical experiments with the ECHAM4/OPYC3 coupled general circulation model (Roeckner et al., 1999) and with the IAP RAS climate model (CM) of intermediate complexity (Petoukhov et al., 1998; Handorf et al., 1999; Mokhov et al., 2000). With the former model two runs are analysed, denoted as GG and GA. In the GG run the model is forced by greenhouse–gases atmospheric loading according to the observations for 1860–1990 and to the scenario IS92a (IPCC, 1992) for 1991–2100. The GA run is forced by the greenhouse–gases atmospheric loading (the same is it was for the GG run) and by the aerosol loading which again is given by the scenario IS92a. In the model both direct and indirect radiative aerosol forcings are incorporated. This scenario ends in 2050. For the IAP RAS CM only the run similar to GG is analysed but here only the effect of CO_2 is taken into account. Forcing due to other greenhouse gases is neglected.

Permafrost cover dynamics is diagnosed using so-called "relative severity index" (Nechaev, 1981) based on surface air temperature. This index is tested against both the present–day cryological observations and paleoclimatic reconstructions (Nechaev, 1981). One sees, however, that only the continuous permafrost cover (when space discontinuity is not higher than 20%) could be fairly reproducible by IAP RAS CM (having the resolution 4.5^o x 6^o) and ECHAM4/OPYC3 (which is run in the resolution T42). A simulation of discontinuous and sporadic permafrost needs higher resolution climate models. As a result most attention here is paid for continuous permafrost cover. Discontinuous one is touched only briefly.

Both models reproduce the present–day permafrost cover fairly well (computed as mean value for 1961– 1990), see Table 1.

Table 1: Present–day permafrost cover, in 10^6km^2, simulated by the models.

	total permafrost cover	continuous permafrost cover
ECHAM4/OPYC3, GG	25.7	12.7
ECHAM4/OPYC3, GA	26.2	13.1
IAP RAS CM, GG	22.3	11.4

During anthropogenically forced scenarios climate warms and permafrost cover degrades. In the run GG the total area covered by permafrost for the both studied models shrinks to about 20 10^6 km^2 in 2050 and to about 17 - 18 million km^2 in 2100. This is solely due to continuous permafrost cover decrease. This type of permafrost amounts about 6 - 7 10^6 km^2 and 2.5 - 3.0 10^6 km^2 in 2050 and 2100, respectively, Fig. 1. In GA smaller (in comparison to GG) warming results in reduced permafrost cover degradation: in 2050 the total permafrost covered area is about 21 million km^2 and the area covered by continuous permafrost is 8 10^6 km^2 Fig. 1. Discontinuous permafrost cover stays practically unchanged during both GG and GA for the both models.

Here permafrost cover sensitivities are formulated in terms of coefficients of linear regressions of normalized (to the simulated present–day value) area covered by permafrost on globally averaged annual mean surface air temperature T_{gl}. In Table 2 the values for this coefficient for continuous permafrost cover are shown. All of them are of the order of 0.3. Corresponding regression plots are presented in Fig. 2. As it was mentioned above, permafrost recession is very similar for the both studied models and the slight dissimilarity in this coefficient is due to the difference in sensitivity of T_{gl} to anthropogenic forcing: in the runs GG ECHAM4/OPYC3 exhibits a higher–magnitude response to greenhouse–gases atmospheric loading than IAP RAS CM. In comparison to 1961–1990, T_{gl} is increased for the former model by about 1.8 K in 2050 and on about 3.4 K in 2100. For the latter these values are only 1.3 K and 2.3 K. This difference is due to the different specifications of forcing in runs GG with two models: (see above). For ECHAM4/OPYC3 the warming proceeds slower in the GA run (than in GG) and reaches about 1.3 K in 2050.

The modelled continuous permafrost cover sensitivity can be compared to the coefficient derived from paleoclimatic reconstructions for two time slices: the Holocene Optimum (HO, 6 ka BP) and the Eemian Interglacial (EI, 125 ka BP). At these time slices T_{gl} was higher than the present day values by 1 K and 2 K, respectively (Velichko and Nechaev, 1992; Velichko et al.,1995). This was accompanied by two– and fivefolds decrease of continuous permafrost cover, correspondingly (Velichko and Nechaev, 1992; Velichko et al.,1995). The points corresponding to these time slices are shown in Fig. 2 together with the respective error bars as crosses. Paleo reconstructed values give sensitivity coefficients ranging between 0.5 K^{-1} (HO–present) and 0.3 K^{-1} (OH–EI). These values are rather similar to those obtained from the model simulations (Fig. 2). Additionally, one may note that changes in T_{gl} simulated by ECHAM4/OPYC3 to 2050 in GG and GA runs are similar to those obtained from the paleco reconstructed data for the Eemian Interglacial and the Holocene Optimum, respectively. An inspection of the corresponding patterns of relative severity indices shows that the modelled and paleo reconstructed patterns are also very similar to each other. It should be emphasised that the above analysis is based on the diagnostic index and the response of the real permafrost cover has to be delayed in comparison to the surface temperature response. The characteristic scale for such a delay is of the order of several tens of years (Velichko et al.,1995).

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Table 2: Coefficient of linear regression of the normalized continuous permafrost cover (with 30–year moving average) on T_{gl}, in K^{-1}, for selected periods.

	1900–2050	1900-2100
ECHAM4/OPYC3, GG	-0.37	-0.27
ECHAM4/OPYC3, GA	-0.30	—
IAP RAS CM, GG	-0.33	-0.34

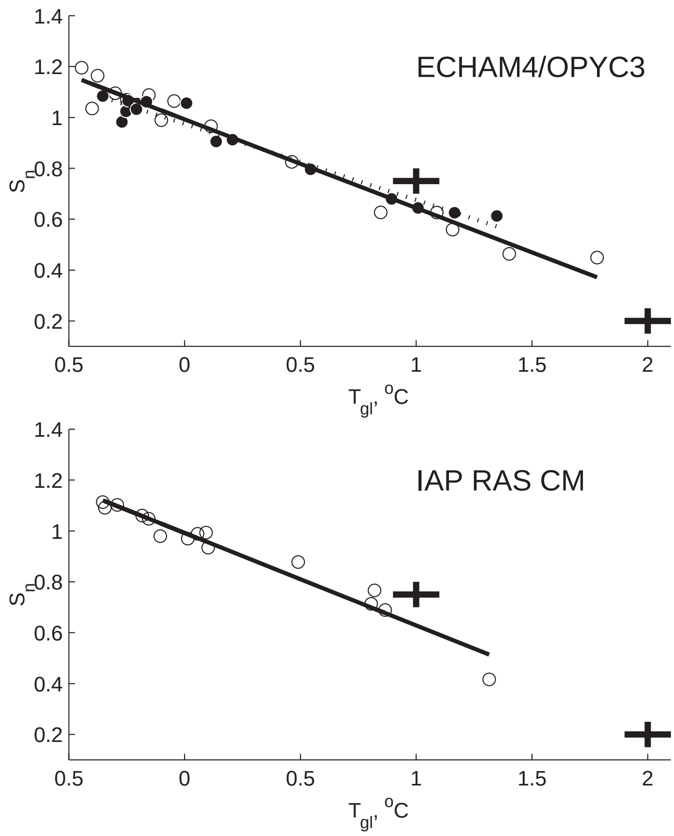

Fig. 2: Dependence of diagnosed area S_n *covered by continuous permafrost on globally averaged annual mean temperature change for ECHAM4/OPYC3 (upper panel) and IAP RAS CM (lower panel). Temperature is depicted in terms of its mean value for 1961-1990. Closed circles present the data from the run GA, open circles - from the run GG. Solid and dotted lines represent corresponding linear regressions. Crosses stand for the data obtained from paleoclimatic reconstructions (together with error bars).*

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ENSO-Monsoon Weakening : Is Global Warming really the Player?

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Introduction

The most significant development in climate research over the recent past has been through the many studies on the global impacts of El Niño Southern Oscillation (ENSO) phenomenon. ENSO is now recognized as the single most important mode of the earth's year-to year climate variability.

During the last two decades, evidence has been accrued for the link between the Indian monsoon rainfall (IMR) and ENSO, first suggested by Sir Gilbert Walker nearly 100 years ago. In general, studies have shown that the warm phase (ie El Niño) of the ENSO phenomenon is associated with the weakening of the Indian monsoon with an overall reduction in rainfall, while the cold phase (i.e. La Niña) is associated with the strengthening of the Indian monsoon with enhancement of rainfall. However during recent times it has been observed that the ENSO – IMR relationship has been weakening. None of the El Niños after 1990 (prolonged 1991-94 and severe 1997) have had any adverse impact on IMR. Several observational/model simulated factors have been attributed to this weakening:

(a) Modulation by the decadal variability of monsoon rainfall (Kripalani and Kulkarni, 1997; Kripalani et al., 2001)

(b) Chaotic nature of monsoon (Webster and Palmer, 1997)

(c) Linkages with the Indian Ocean Dipole Mode (Saji et al., 1999; Behera et al., 1999; Webster et al., 1999)

(d) Global Warming (Krishna Kumar et al., 1999; Ashrit et al., 2001)

(e) Atlantic Circulation (Chang et al., 2001)

Chang et al. (2001) suggested that the weakening of the ENSO-IMR relationship may be due to the strengthening positive phase of the North Atlantic Oscillation (NAO). This positive phase of NAO is again attributed to global warming.

If global warming is really the cause of the recent weakening , the relationship between the Northern Hemisphere Surface Air Temperature (NHST) and Eurasian snow cover (as indicators of global warming) with IMR should strengthen in recent times. This article examines this realtionship.

Data and Methodology

(i) The most effective parameter in explaining IMR variability compared to other indices of the ENSO phenomenon is the Darwin Pressure Tendency DPT:April minus January (Shukla and Mooley, 1987). Data are used for 1871 – 2000. The relationship of DPT with IMR is negative.

(ii) NHST anomalies – average for January and February. Data are used for 1871-2000. The relationship of NHST with IMR is positive.

(iii) Satellite-derived snow estimates are available only since mid 1960s. However in a recent study Kripalani and Kulkarni (1999) have shown that January snow depth over western (eastern) Eurasia is negatively (positively) related with subsequent IMR. This study used the observed historical Soviet snow depth data product developed at the National Snow and Ice Data Center, USA under the bilateral data exchange agreement between USA and former USSR. The data period for Version II of this dataset varies from 1881 to 1995. The length of data period over western Eurasia is sufficiently long. Hence a time series of snow depth for the month of January over this region has been prepared for the period 1891-1995. Snow depth over this region and IMR are negatively related.

(iv) IMR series for the period 1871-2000 available on the web site of the Indian Institute of Tropical Meteorology: http://www.tropmet.res.in

To examine how the relationship of IMR with DPT, NHST and snow have varied with time on decadal basis, 11-year sliding correlation coefficients (CC) are computed. With longer window lengths of 21-year or 31-year period information at the end points will be lost in particular during the recent past decade. The decrease in magnitude of the relationship between DPT and IMR should signify the weakening over the Pacific, while the increase in magnitude with NHST and snow should signify strengthening over the Northern Hemisphere / Eurasian continent, suggesting possible role of global warming in these changing relationships. The significant value of correlation coefficient for a sample of 11 at 5% significance level is ~0.57.

Results

Figure 1 (top panel, page 17) shows the 11-year sliding CCs of DPT with IMR. The relationship appear to be weak up to 1930. However during the period 1930-1970 the relationship amplifies in magnitude and thereafter starts weakening with CCs near zero in early 1990s. This clearly depicts the recent ENSO – IMR weakening.

Figure 1 (central panel) shows the CCs of NHST with IMR. Again till 1930 the relationship fluctuates. As with DPT, the relationship increases during the 1930-1970 period and drops drastically thereafter changing sign around 1990.

Figure 1 (bottom panel) shows the CCs of snow with IMR. The relationship increases in magnitude during the 1950-1980 period and decreases thereafter with CCs near zero around 1990. Thus the weakening of IMR with DPT, NHST and snow appears to have commenced around the same time ie 1970, with CCs near zero and changing signs around 1990.

Chang et al. (2001) have noted that the correlation between IMR and winter western and central Eurasian temperature has become significantly positive in recent decades using a 21-year sliding window. With this window length the information for the last 10-year period (1989-1998) used by them is lost. A careful examination of Figure 1 (CLIVAR Exchanges, No 20, June 2001, pg 1) from Chang et al. clearly shows the correlations dropping since early 1980s. Hence a smaller window length of 11-year period may give more information at the end points in particular after 1990s.

Thus the drastic fall in CCs of IMR with NHST and snow in recent times do not support the view that global warming is a cause of recent ENSO-Monsoon weakening. Such weak relationships have been noted earlier eg DPT during 1920-1930; NHST during 1930-1940 and snow prior to 1950. Further the relationships of IMR with several of its regional and global predictors have undergone secular variations is well documented.

Summary

Analysis of more than 100 years of observed data confirm the recent ENSO-Monsoon weakening. However the weakening of IMR-NHST and IMR-Snow relationships in recent times do not support the global warming as a possible cause of changes in ENSO-Monsoon links. Such changes have been observed in earlier times.

Our recent study (Kripalani and Kulkarni, 2001) has shown that monsoon related events (rainfall over South Asia, rainfall over East Asia, circulation over Pacific, circulation over northern hemisphere) over geographically separated regions seem to get linked (or de-linked) around the same time. Around 1990 IMR appears to have de-linked not only with the Pacific but also with Northern Hemi-

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Banks and Wood: Can coupled models help to define an observing strategy for detection of climate change in the ocean?, page 7:

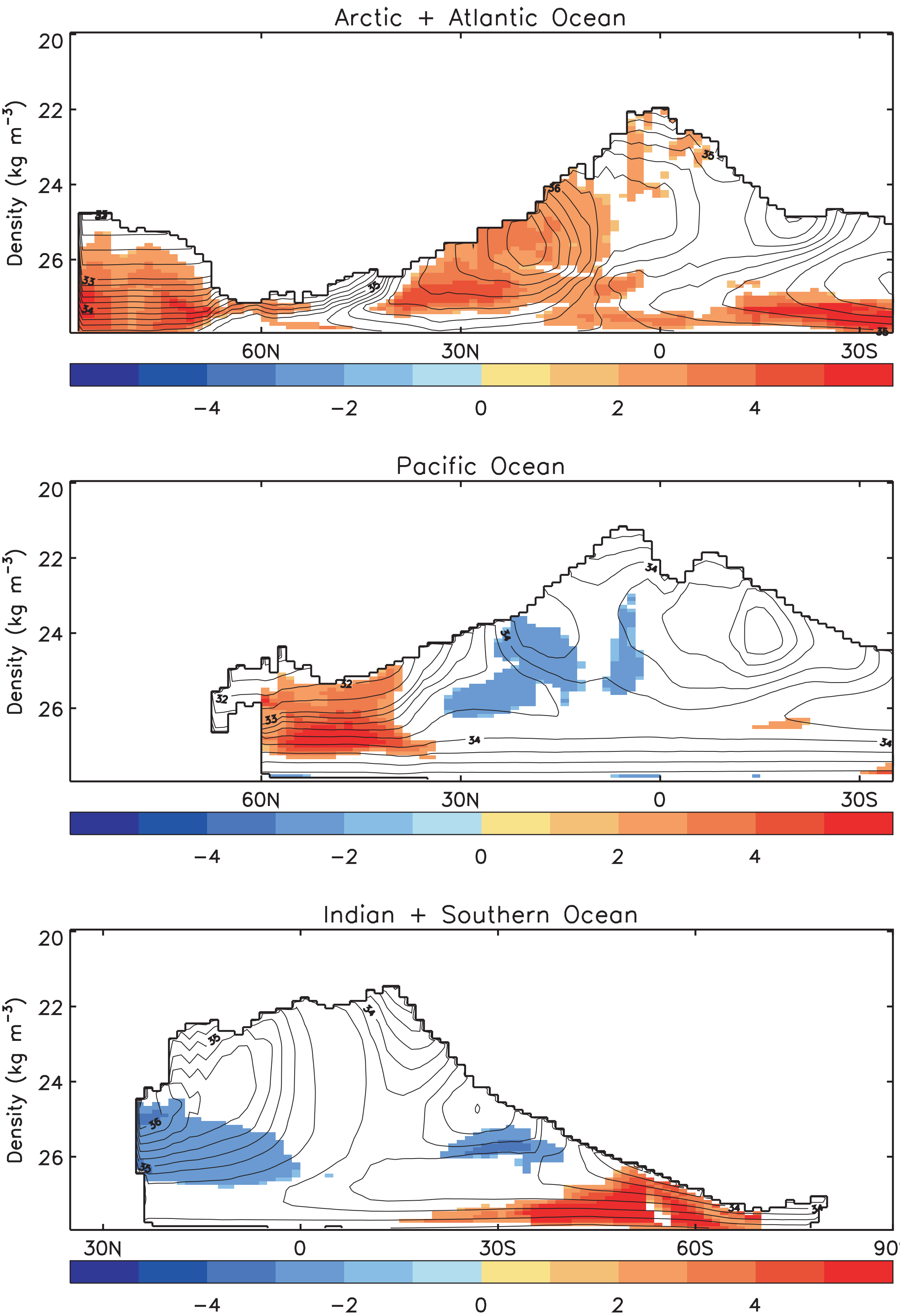
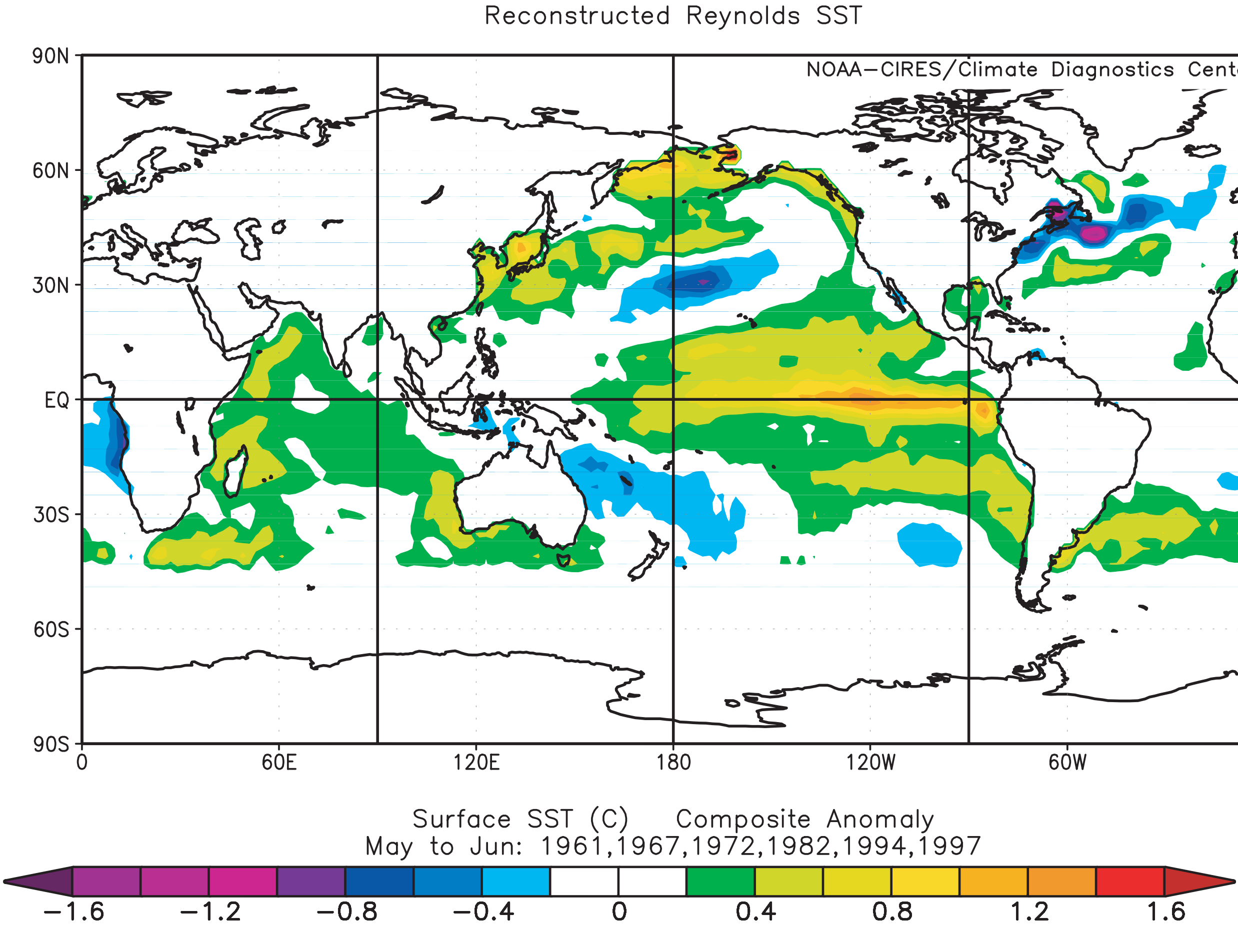
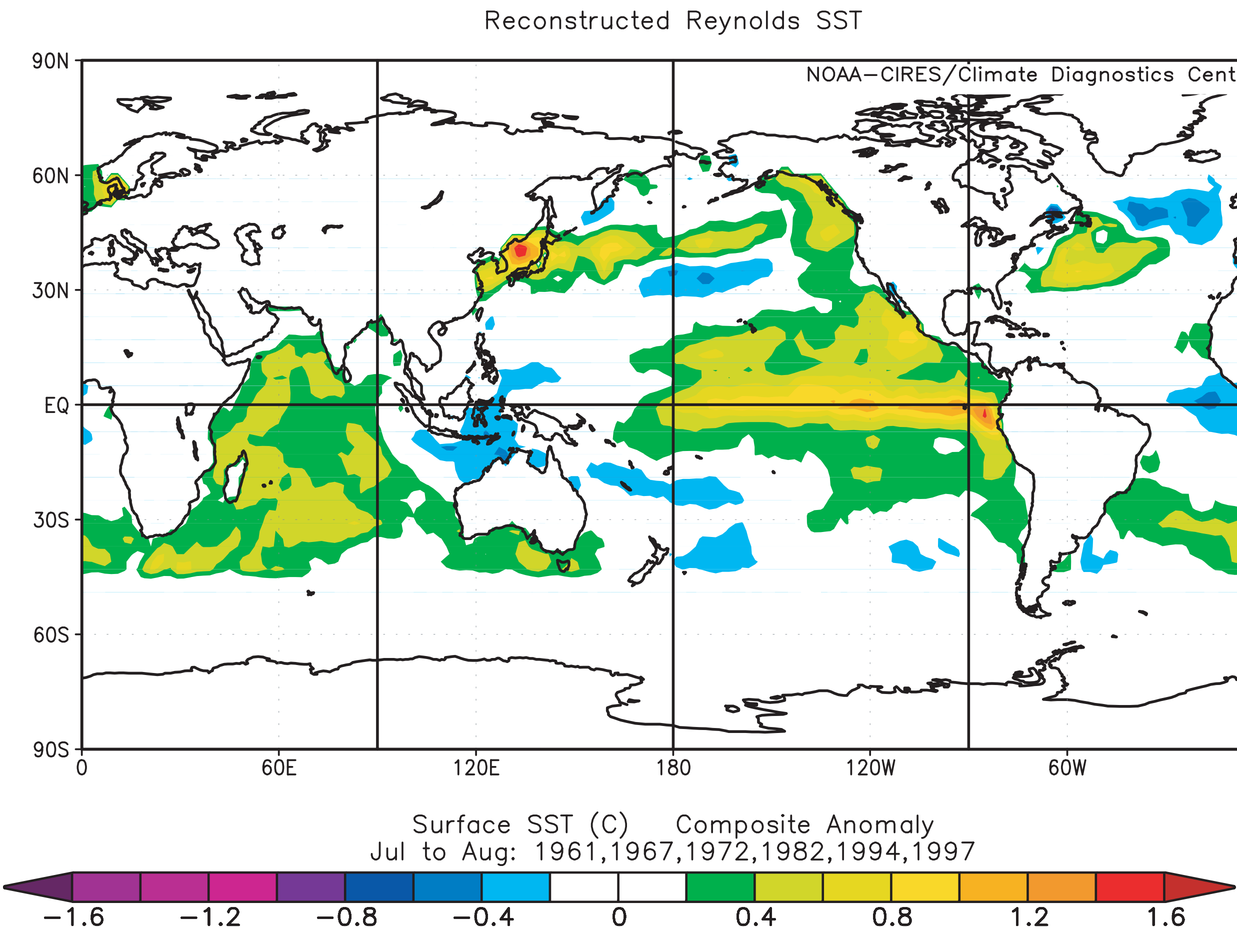


Figure 2: Zonally averaged signal-to-noise ratios for salinity on isopycnals for each basin (shown in colour only when significant at the 5% level). The contours indicate the background salinity field and enable water masses to be identified. The modelled water masses are qualitatively similar in structure and distribution to observed water masses. SAMW can be seen around 26 kg m^{-3} and 30^{o}S in the Indian Ocean.

Allan et al.: Is there an Indian Ocean dipole, and is it independent of the El Niño - Southern Oscillation?, page 18:





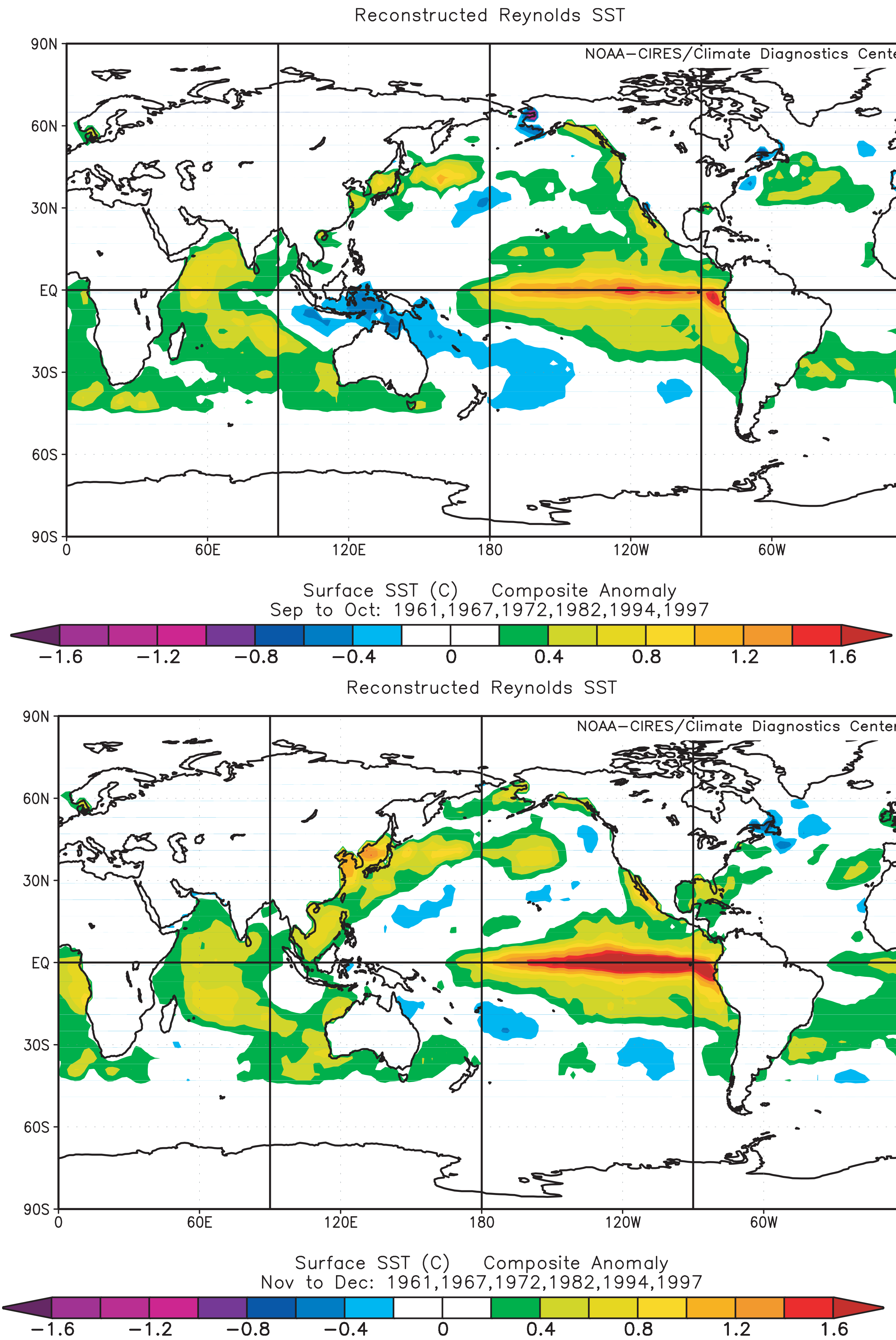


Figure 4: Composite of years and months in the sequence from Saji et al. (1999) extended over the full Indo-Pacific domain for SST anomalies.

Yasunaka and Hanawa: Regime Shifts in the 20th Century Found in the Northern Hemisphere SST Field, page 22:

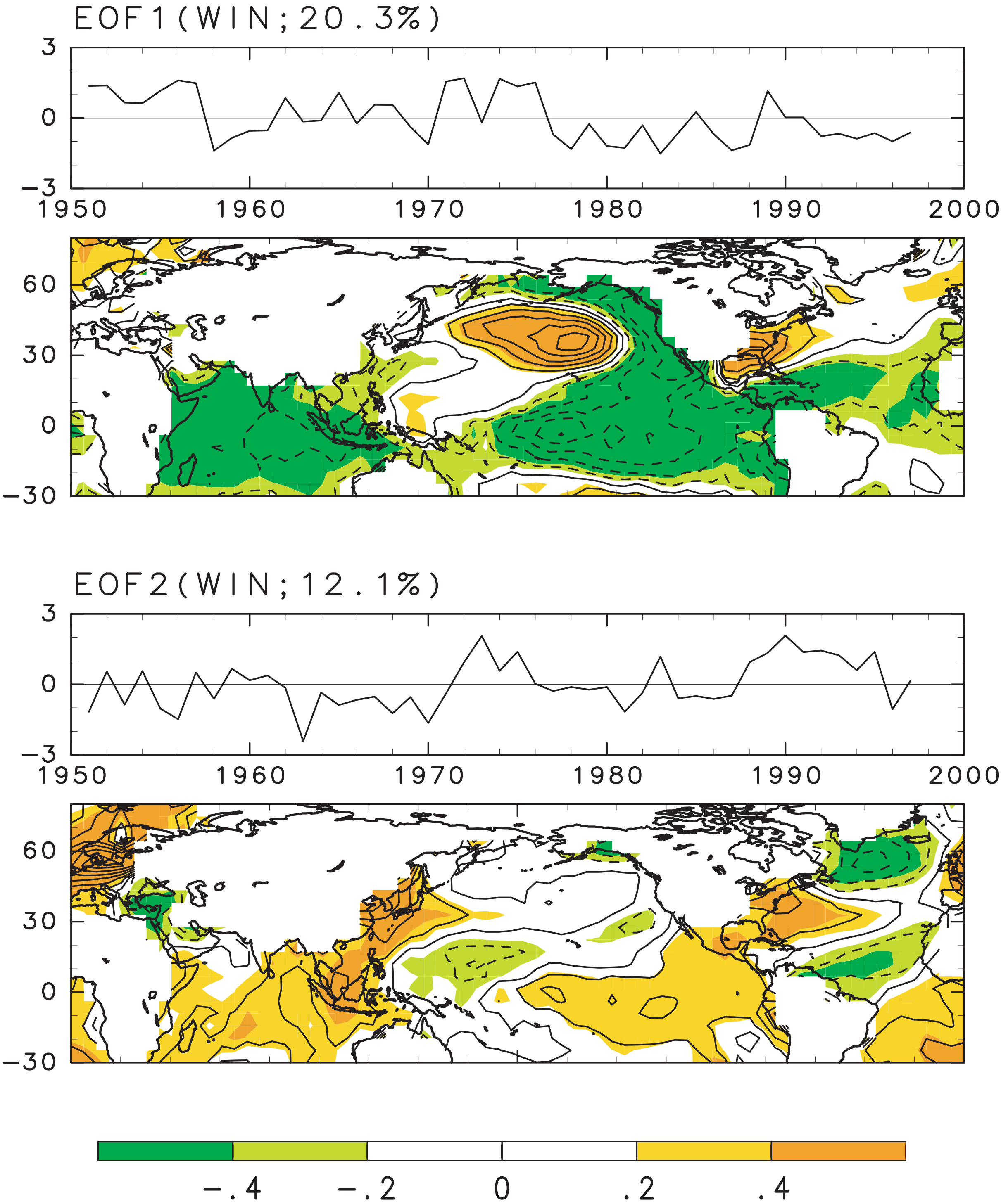


Fig. 1: Standardized time coefficients of the first and second EOF modes, for the Northern Hemisphere winter (January through March) mean SST anomalies from 1951 to 1997, and regression maps of the winter mean SST. Contour interval is 0.1°C. Shadings are given based on correlation coefficients with the time coefficient. Negative contours are dashed.

Kripalani et al.: ENSO-Monsoon Weakening : Is Global Warming really the Player?, page 11:

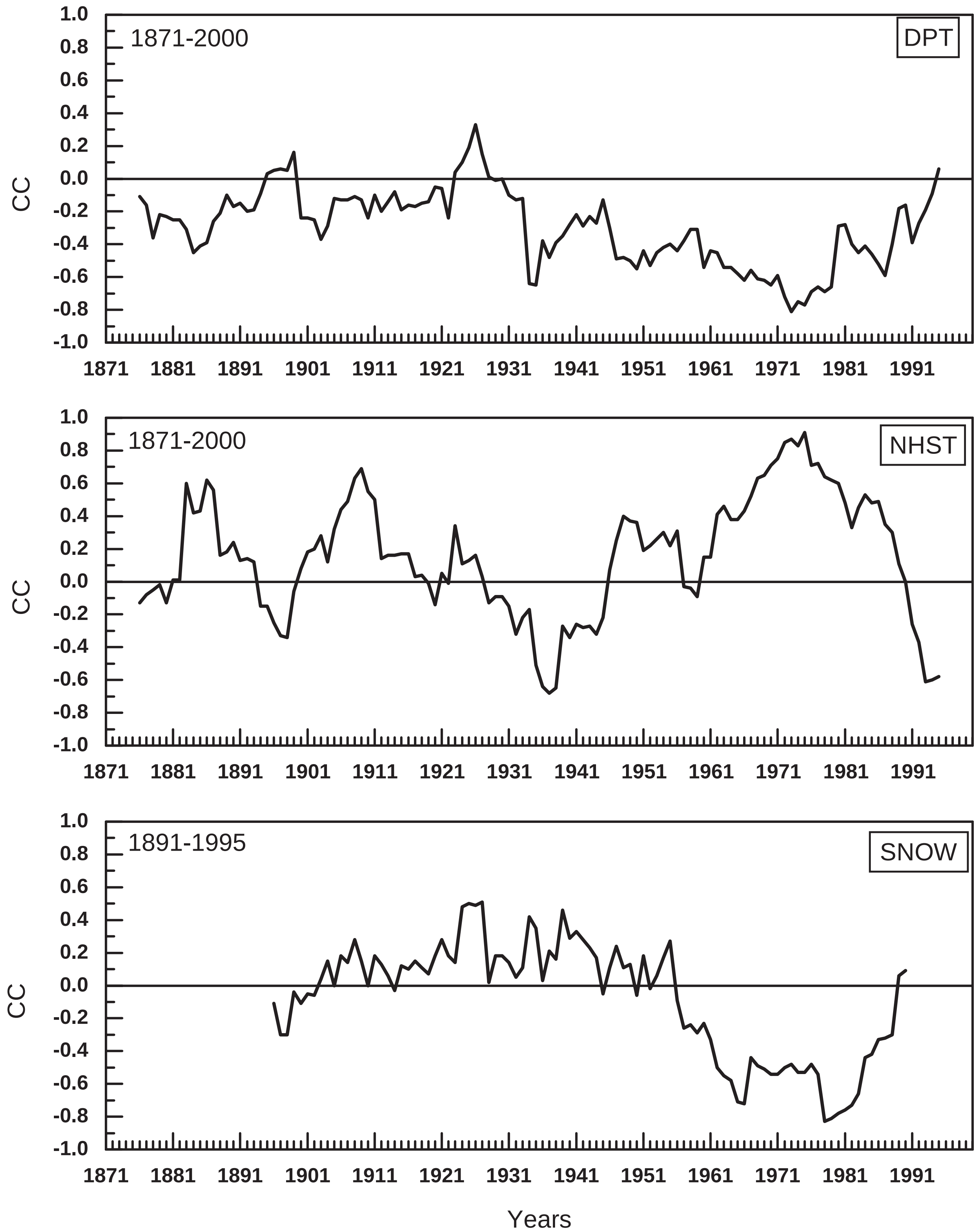


Fig. 1: 11-year sliding correlation of IMR with DPT, NHST and Snow. The values are plotted at the centre of the 11-year period.

continued from page 12

sphere/Eurasian continent. Hence the recent ENSO-monsoon changes may be just a part of natural climate variability.

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Is there an Indian Ocean dipole, and is it independent of the El Niño - Southern Oscillation?

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The papers by Saji et al. (1999) and Webster et al. (1999) describe an equatorial Indian Ocean sea surface temperature (SST) dipole pattern (IOD) which they claim modulates rainfall in East Africa and Indonesia, and operates independently of the global-scale El Niño - Southern Oscillation (ENSO) (Tourre and White, 1997). The concept of possible independence of Indian Ocean SST variability from ENSO has been shaped by research focusing on climatic events during the 1990s (eg. Behera et al., 1999; Murtugudde et al., 2000). However, in an earlier paper Nicholls (1989) describes a different IOD pattern of variability related to Australian winter rainfall, and argues that this pattern operates largely independently of ENSO. In this paper we demonstrate clearly that with consideration of the evolution of ENSO events, the varying lag correlations between IOD and ENSO indices, and using seasonally-stratified data, the apparent ENSO independence disappears from both IODs.

IOD SST patterns and indices were identified by Saji et al. (1999) and Webster et al. (1999) using a mixture of Empirical Orthogonal Function (EOF) and correlation techniques, oceanic and atmospheric observations and dynamics. However, Saji et al. (1999) fail to take account of the spatiotemporal evolution of ENSO events when using a standard EOF analysis of data in a non-temporally stratified form. They mistakenly assume that the EOF 1 and 2 modes in their results are distinct phenomena, being ENSO and an independent IOD signal. Although these EOFs are orthogonal at zero lag, they are confounded and have considerable shared variance as their time series correlate significantly with one another at leads and lags of around 9-10 months. Such problems have been highlighted further using EOF Varimax rotation (Tourre and White, 1995) and in two recent papers focusing on various EOF (non-rotated and rotated) and Principal Oscillation Pattern (POP) examinations of the IOD question (Dommenget and Latif, 2001; Baquero-Bernal and Latif, 2001). These studies indicate that the two nodes of the IOD are not significantly anti-correlated in time, and the IOD structure is part of ENSO. The lack of consistent anti-correlation between eastern and western IOD nodes is also seen when the data are detrended and ENSO influences are removed (Nicholls and

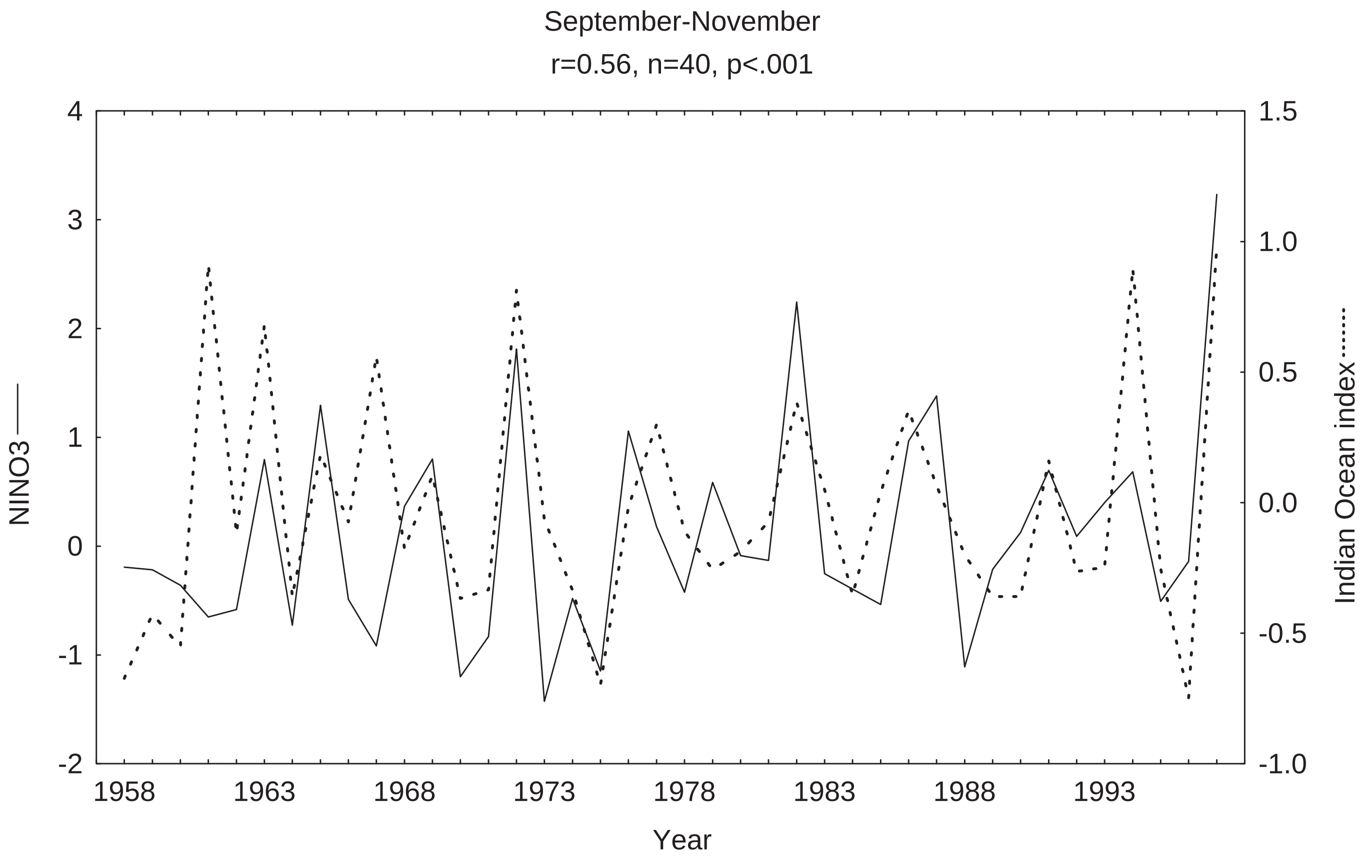

Figure 1: September-November Saji et al. (1999) IOD and Niño 3 SST indices, post 1957 (r=+0.56).

Drosdowsky, 2001). Nevertheless, the above EOF and correlation mistakes are still repeated in recent IOD papers (Behera et al., 2000; Behera and Yamagata, 2001; Iizuka et al., 2000).

Both Saji et al. (1999) and Webster et al. (1999) report insignificant correlations between their IODs and the Niño 3 region (5°N-5°S, 150°W-90°W) SST index of ENSO. A different picture emerges, however, if the correlations are calculated on monthly or seasonally-stratified values of the indices. Thus, the correlation between mean September-November values of the Saji et al. (1999) IOD index and Niño 3 is 0.52, using data from 1872-1997. The correlation using only the shorter post-1957 period examined by Saji et al. (1999) is 0.56.

Both of these correlations are highly statistically significant. Replacing the Saji et al. (1999) IOD index with the slightly different Webster et al. (1999) IOD index produces correlations only marginally different (and still very significant). Correlations between the Saji et al. (1999) IOD index and Niño 3 using March-May means are, however, weakly negative (and not statistically significant). These findings are confirmed in a similar correlation analysis using monthly-stratified data (Nicholls and Drosdowsky, 2001). An important aspect of this work is that removal of the long-term trend and the ENSO signal still fails to reveal distinct evidence of a regular IOD (Figure 2). A similar situation is observed across the Indian Ocean sector when the ENSO signal related to the Southern Oscillation Index (SOI) is removed during the peak of IOD activity (Mutai and Ward, 2000). The situation may be more akin to the Atlantic Ocean, where the meridional SST anomaly gradient, and not a dipole, is the important physical feature (Rajagopalan et al., 1998).

Webster et al. (1999) assert that the intensity of the dipole is independent of the very strong El Niño event during 1997. September-November values of the IOD, Niño 3 SSTs, the SOI and Darwin mean sea level pressure (MSLP) do not support this conclusion (Figures 1 and 3). The strength of the 1997 IOD index is consistent with the very strong 1997 El Niño. There are years (1961, 1967, 1994) when the IOD intensity is substantially diffeent from what could be expected given the Niño 3 index for that year (Figure 1). However, the 1961 and 1994 IOD events do show responses to the SOI and Darwin MSLP (Figure 3), and occur during 'protracted' La Niña and El Niño episodes respectively (Allan and D'Arrigo, 1999; Allan, 2000; Reason et al., 2000; Allan et al., 2001), when an IOD pattern is found in conjunction with enhanced Niño 3.4 to 4 SST anomalies. The 1967 IOD develops during the onset of a La Niña event, as noted by Saji et al. (1999).

Other evidence supports the interdependence of ENSO and the IOD. Graham and Goddard (1999) find a peak correlation of 0.75 at a 3 month lead between Niño 3.4 (5°N-5°S, 170°W-110°W) SSTs and a central equatorial Indian Ocean SST index (0-15°S, 50°E-80°E) of the western node of the IOD. Mutai et al. (1998) show highly signifi-

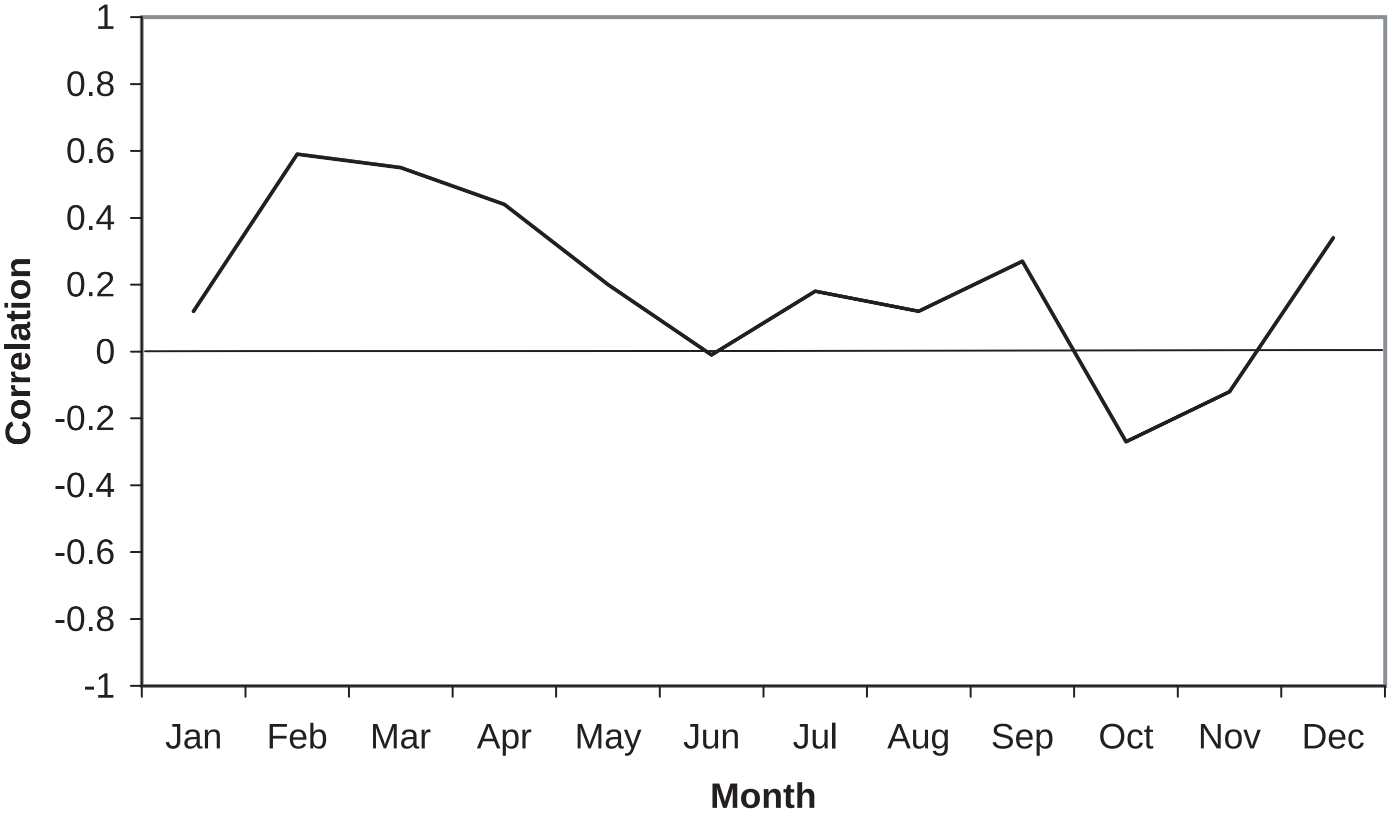

Figure 2: Correlations between east and west equatorial Indian Ocean SST, after detrending and removal of ENSO influence (Niño 3 effect removed).

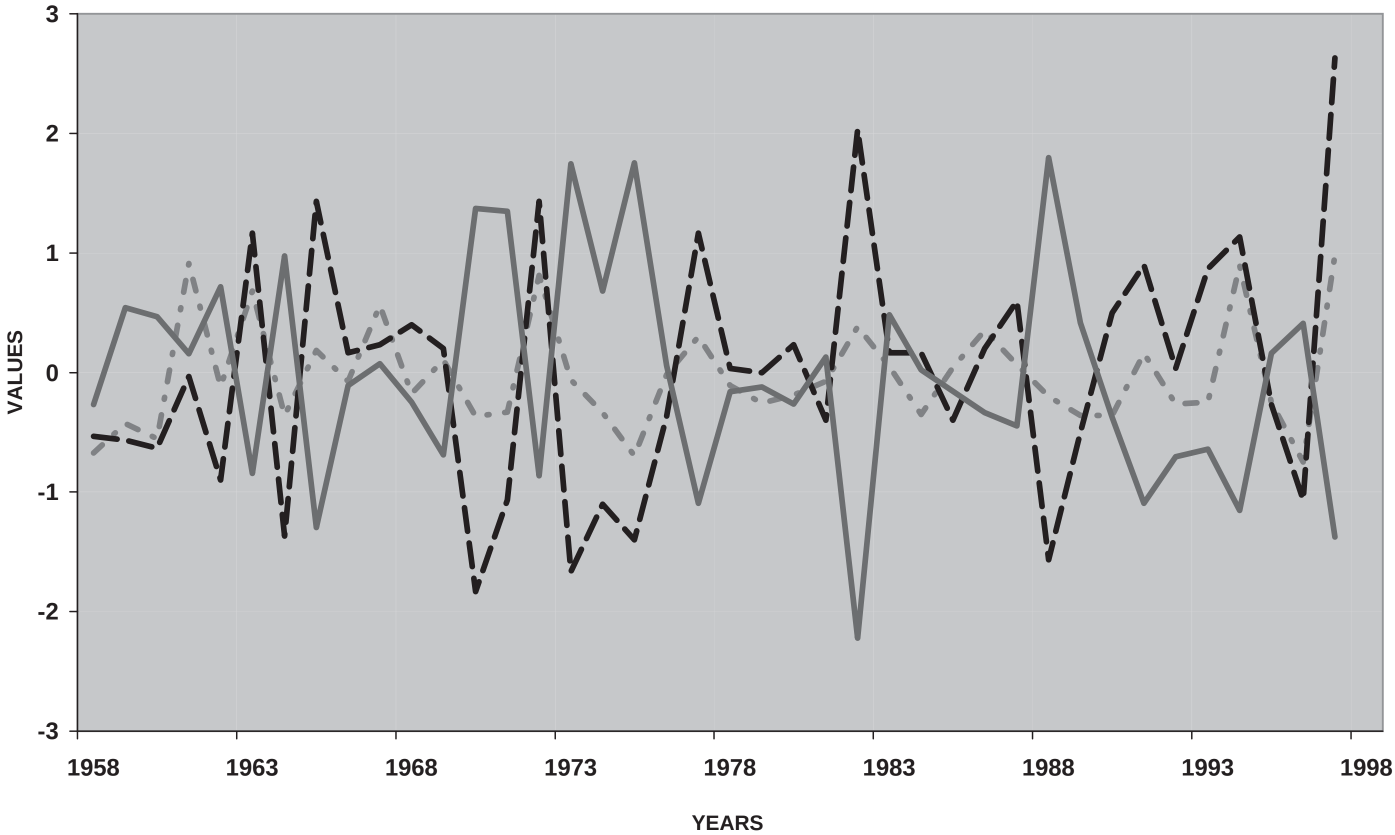

Figure 3: September-November Saji et al. (1999) IOD (dashed dotted) versus the SOI (divided by 10, solid, r=-0.56) and Darwin MSLP (dashed, r=+0.69), 1958-1997.

cant correlations between July-September Niño 3.4 and Niño 4 (5^{o}N-5^{o}S, 160^{o}E-150^{o}W) SSTs and October-December East African rainfall occurring in conjunction with an IOD. Chambers et al. (1999), Stone et al. (1996) and Wright et al. (1985) provide evidence that, taken together, supports significant relationships between ENSO and East African rainfall during IOD events. Reason et al. (2000) and Allan et al. (2001) have analysed composites of the evolution of atmospheric and oceanic variables over the Indian Ocean for both strong ENSO events and 'protracted' ENSO episodes. IOD patterns are again clearly evident during the evolution of both of these ENSO types. In fact, the IOD composites of Saji et al. (1999) have been shown to be synchonous with El Niño events when examined across the full Indo-Pacific basin (Hendon, 2000, per. com.). Such ENSO-IOD structures can also be produced (Figure 4, page 14/15) on the NOAA WWW site (http://www.cdc.noaa.gov/Composites/). Velocity potential anomalies for the IOD composites in Figure 4 show simply the rearrangement of the Walker Circulation across the Indo-Pacific basin that occurs during the evolution of El Niño events (Hobbs et al., 1998).

Saji et al. (1999) and Webster et al. (1999) propose local coupled mechanisms to explain the IOD. However, Chambers et al. (1999) explain the IOD by local wind forcing that is strongly correlated to the SOI. An IOD-like structure in SST anomalies is explored by Baquero-Bernal and Latif (2001) using a control integration with a coupled ocean-atmosphere general circulation model (GCM), a coupled run in which ENSO is suppressed, and one run in which the ocean GCM is replaced by a mixed layer model. These experiments quantify the contributions of ENSO and ocean dynamics to SST variability in the tropical Indian Ocean. The results show that ocean dynamics are not important to this type of IOD-like SST variability. It is forced by surface heat flux anomalies integrated by the thermal inertia of the oceanic mixed layer, which reddens the SST spectrum. These experiments find no evidence for an IOD generated by ocean dynamics that is independent of ENSO. However, several other GCM studies (Behera et al., 2000; Iizuka et al., 2000; Vinayachandran et al., 2001) continue to interpret and analyse their results based on the IOD concept detailed by Saji et al. (1999), or extended by Behera and Yamagata (2001). Although other recent research has somewhat questioned the IOD independence of ENSO (e.g., Li and Mu, 2001; Mu and Li, 2001), the underlying assumption that an independent IOD mode exists still continues to be perpetuated.

This note is not intended to suggest that no other climatically driven signals exist in Indian Ocean SSTs other than those attributed to ENSO. However, it does indicate that the IOD feature reported in the literature is an integral part of ENSO evolution. Thus, every effort must be made to take account of the full ENSO-related climatic signal in analyses seeking to investigate other phenomena in the Indian Ocean domain.

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Regime Shifts in the 20th Century Found in the Northern Hemisphere SST Field

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1. Introduction

A 'regime shift' is characterized by an abrupt transition from one quasi-steady climatic state to another, and its transition period is much shorter than the lengths of the individual epochs of each climatic state. In the present study, we investigate when regime shifts occurred and what was the difference in climatic states before and after the shifts, using the wintertime sea surface temperature (SST) field in the Northern Hemisphere. The relationship between changes in the SST field and those in the atmospheric circulation is also investigated. Complete details of our results are given by Yasunaka and Hanawa (2001).

2. Dominant variations of the winter Northern Hemisphere SST field

First, in order to detect organized patterns of the SST variations in the Northern Hemisphere, we adopted an empirical orthogonal function (EOF) analysis. Figure 1 (page 16) shows the standardized time coefficients of the first two EOF modes, and the distributions of regression and correlation coefficients of the winter mean SST anomalies with the standardized time coefficients of EOF modes. It is found that the dominant modes correspond well to the specific atmospheric circulation patterns. That is, the first mode resembles the response to the activity of the Pacific/North American (PNA) pattern, and is similar to SST changes in El Niño/Southern Oscillation (ENSO) and so-called Pacific Decadal Oscillation (PDO). The second mode has high correlation with the activity of the Arctic Oscillation (AO). EOF analyses to each oceanic basin separately are also made and the robustness of these modes has been confirmed.

3. Detection of regime shifts

In order to identify the years when regime shifts occurred in the SST field, we carefully inspected the time series of original gridded SST data and those of the EOF modes. If the years of the regime shifts detected by both of the two time series are same, then we can say that the significant and systematic regime shifts occurred in these years. The difference between the 5-year means before and after the given year is used as a measure of shifts. As a result, seven regime shifts were detected during the period from the 1910s to the 1990s: 1914/15, 1925/26, 1945/46, 1957/58, 1970/71, 1976/77 and 1988/89, as shown by arrows in Fig. 2. It was also found that all these regime shifts could be found in both or one of the first and the second SST-EOF modes.

Figure 3 shows the SST difference between the 1971-76 regime and the 1977-88 one, that is, the periods before and after the 1976/77 regime shift. It reveals the changes in an intensity of the Aleutian Low (AL) and the corre-

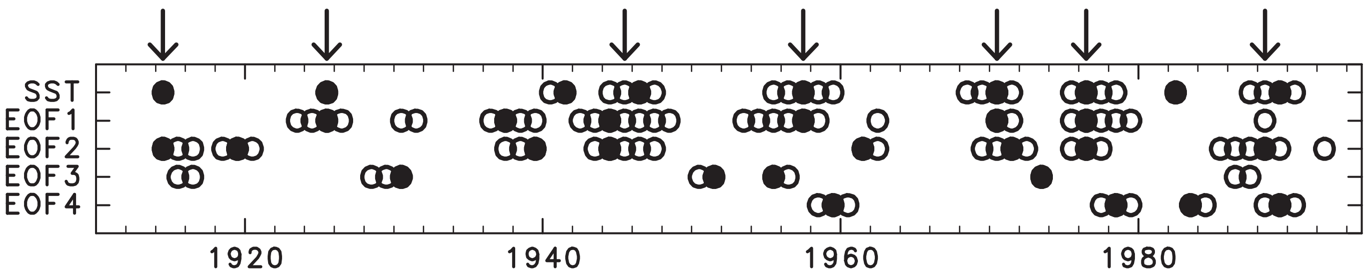

Fig. 2: Years when significant shifts occurred in the SST field and the first four EOF modes. Closed circles denote the year showing the maximum shift in a cluster of significant shift years (open circles). The years shown by arrows are those designated as regime shifts at the present study.

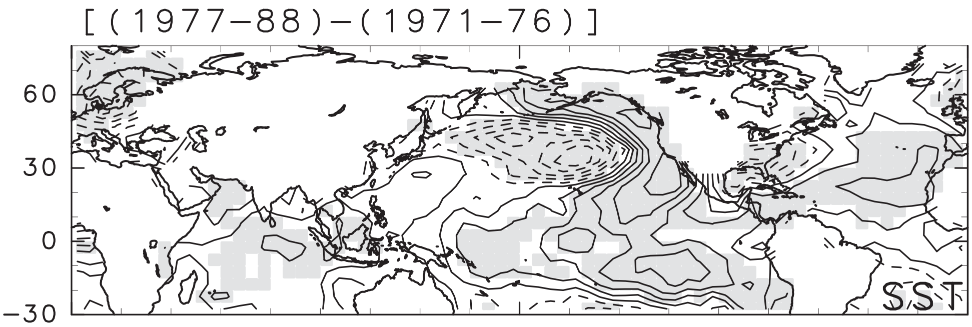

Fig. 3: Map showing difference in winter means of SST between the two periods of 1971-76 and 1977-88. Contour interval is 0.2℃. Negative contours are dashed. Dotted regions are those in which the differences exceed 90% significant level.

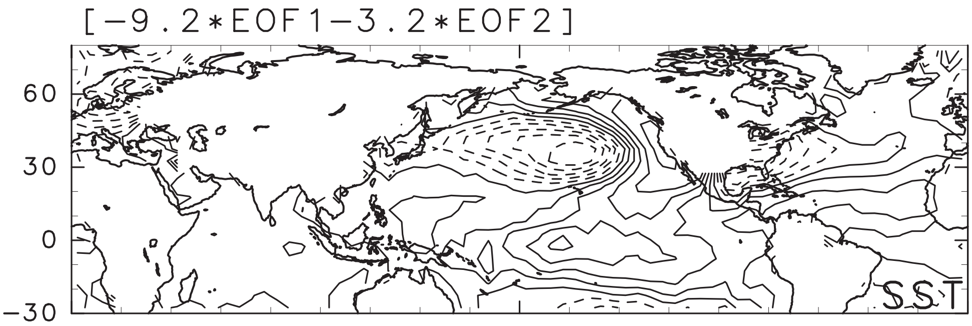

Fig. 4: Reconstructed map using the first and second SST-EOF regression maps of SST, which are summed with weights according to their magnitudes in the 1976/77 regime shift. Contour interval is 0.2℃. Negative contours are dashed.

sponding SST changes in the central North Pacific. This regime shift represented as the step function can account for 29.1% of the total variance in the SST field for 18 years from 1971 to 1988.

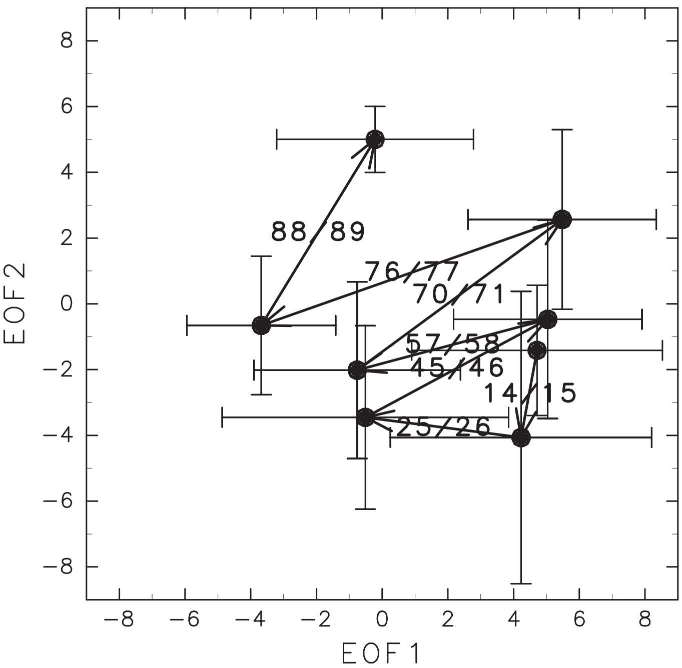

Fig. 5: Trajectory of the regime shifts on the diagram of the first and the second EOF time coefficients. Closed marks with vertical and horizontal bars denote the mean values between the regime shifts with the standard deviations.

Here we try to reconstruct the regime shifts, using the regression maps of the first and the second EOF modes (Fig. 1). Reconstructed maps are made by the superposition of the regression maps with weights according to the shift magnitudes in their EOF time coefficients. Figure 4 is the reconstructed SST difference for the 1976/77 regime shift (-0.92 x EOF1 + (-3.2) x EOF2). The features of these maps are quite similar to those of the difference maps shown in Fig. 3. Actually, spatial correlation coefficient between the difference map (Fig. 3) and the reconstructed map (Fig. 4) is 0.90. It is found that the other regime shifts are also represented well by the reconstructed maps of the first and the second EOF modes as listed in Table 1.

Figure 5 shows a trajectory of the regime shifts represented by the time coefficients of the first and the second EOF modes. Almost all regime shifts except for the 1914/15 shift tend to direct toward the same polarity in both the first and the second EOF modes. This means that the changes of the AL and SST in the North Pacific are enhanced by superposition of the two modes. Six regime shifts except for the 1914/15 shift are significant at 90% level in the first mode and/or the second mode.

Further, we tried to reconstruct each of the regime shifts using the atmospheric indices instead of the EOF modes, i.e. the PNA index (Barnston and Livezey, 1985) and the AO index (Thompson and Wallace, 2000). The results show that the regime shifts are reconstructed very well as shown in Table 1.

4. Conclusion

In the present study, significant six regime shifts except for the 1914/15 regime shift were detected in the Northern Hemisphere SST field during the period from the 1910s to the 1990s. It was found that the first and the second SST-EOF modes, which correspond to the changes in activities of the PNA and the AO, could represent well all these regime shifts. This means that regime shifts occur as superposition of several unique modes (the first and the second EOF modes) of variabilities existing in the climate system. That is, the regime shifts are transitions from one quasi-steady climatic regime to another. The spatial patterns of the regime shifts are not perfectly identical with each other, but have similar feature containing a change of the AL intensity.

Year	Percentage of variance	Coefficients EOF1	EOF2	Pattern correlation	Coefficients PNA	AO	Pattern correlation
1914/15	13.1	0.5	-2.7	0.29			
1925/26	13.8	**-4.7**	0.8	0.47			
1945/46	14.6	**5.5**	**3.0**	0.70			
1957/58	15.2	**-5.8**	-1.5	0.77			
1970/71	21.6	**6.2**	**4.6**	0.86	-0.4	**1.2**	0.75
1976/77	26.8	**-9.2**	**-3.2**	0.93	**1.3**	**-1.0**	0.82
1988/89	19.2	**3.5**	**5.7**	0.80	-0.7	**2.2**	0.86

Table 1: Reconstruction of regime shifts using the two leading EOF modes, and the PNA and the AO indices. Bold numerals show correlation coefficients exceeding the 90% significant level by the Student t-test. Percentage of variance is the ratio of variance accounted for by the shift between the two regimes in the whole variance for the period of two regimes. Pattern correlation is the correlation coefficient between the difference map as shown in Fig. 3 and the reconstructed map as shown in Fig. 4. The 1976/77 means the regime shift between 1971-76 and 1977-88 and so on.

The duration between each regime shift is about 10 years, which are identical to the PDO. In addition, the change in the AL activity, which plays an important role in the PDO due to the tropical forcing through the atmospheric bridge (Gu and Philander, 1997), or due to the local mid-latitude SST forcing (Latif and Barnett, 1996), is included. These imply the existence of some relationship between the regime shifts and the PDO.

In the present study, we showed that abrupt jumps in the second SST-EOF mode reflecting the activity of the AO occurred simultaneously with jumps in the first mode. The AO is known to be the variation of the SLP field coupled with the variation of the polar jet in the stratosphere (Thompson and Wallace, 1998). Although the cause of the AO has not been clarified, Honda et al. (2001) suggested that the AL activity in early through mid winter could influence that of the AO. Therefore, the change in the AL activity associated with the PNA pattern might have some connection with that of the AO. We might speculate that the AL is an important role in the regime shifts.

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Virtual Center for Decadal Climate Variability Studies on the Web

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The first phase of the Virtual Center for Decadal Climate Variability Studies has been developed, to foster speedier communication within the global, decadal climate variability and applications communities; to serve as a facility to access long-term data sets of various types, integrated with analysis and visualization software; and to allow community-wide planning and execution of data set intercomparisons, multi-model experiments, and observing-system studies. The Virtual Center aims also to facilitate closer involvement of these two communities in the planning and implementation of projects such as CLIVAR, GEWEX, and IGBP.

The first phase of this Virtual Center is now ready and everybody interested is invited to become a member of this Center by filling out a membership form on the Help Desk at http://www.DecVar.org. Members will be supplied with a username-password and helpful instructions about how to use this facility. An informative tour and some areas of the Virtual Center can be accessed without becoming a member. Even if not possible to implement it immediately, members' and non-members' feedback will be very useful for subsequent revisions, enhancements, and additions.

The idea of such a Center has been initiated and developed by Dr. Vikram Mehta (Univ. of Maryland-NASA/ GSFC) with steady encouragement and support by Dr. Eric Lindstrom, the NASA-Oceanography Program Scientist. Many months of work by Bob Heinmiller, Fred McDavid, Susan Kubany, Alan Richmond, and Stan Fuller at Omnet, Inc.; and Casey Rexrode and Brian Davis at Night Light Design have gone into taking the concept of this Virtual Center to the first phase of implementation. Andrea McCurdy's suggestions have significantly improved the "look-and-feel" of the Virtual Center layout. Suggestions from Drs. Niklas Schneider, David Pierce, Jim O'Brien, Jerry Meehl, Eric Lindstrom, Yi Chao, and Eileen Shea have significantly improved the ease of using this facility.

One of the unique features of the Virtual Center Website is its interactivity. There are several locations on the Virtual Center Website where a member can upload files of various types (papers, reports, pictures, etc.) and post messages on the various bulletin boards.

The Virtual Center Website has a unique facility, developed by Omnet, for members to edit one or more documents simultaneously. The software for this facility automatically collects and organizes members' comments to make the final editing much easier. Also, each document can be individually password-protected, so only the members authorised to read and edit it can do so. A sample document has been posted for members to practice upon. Members may post unpublished or soon-to-be published work on this Virtual Center Website. It is expected as a matter of decency and courtesy that proper acknowledgement should be made if such information is used by others.

In addition to these interactive areas, there are other areas where members can not post information themselves. There is a large amount of information already posted in some of these areas. Some areas are under construction and will be completed soon. Updates, additions and corrections are very much appreciated and should be sent to Dr. Vikram Mehta (mehta@eos913.gsfc.nasa.gov).

The Virtual Center will publish a quarterly newsletter "Subtle Signals" about various aspects of decadal variability, including societal impacts. The first issue of "Subtle Signals" is available on the Virtual Center Website. Size and format suggestions for newsletter articles are mentioned in the first issue

As mentioned earlier, one of the purposes of this Virtual Center is to facilitate planning and execution of collaborative projects on various aspects of decadal variability. In preliminary discussions, scientists from many organizations have shown a strong interest in such projects. Discussion among the interested scientists about collaborative projects will soon been started. The next and subsequent phases of development of the Virtual Center will consist of easier-to-find organization of information within each area, search facilities in many areas, discussion sessions-poster sessions-seminars-conferences, long-term data sets of various types, analysis and visualization software, and much more. People interested in becoming involved in this Center's activities, including collaborative projects, should contact Dr. Vikram Mehta.

The Virtual Center belongs to the decadal variability community. It is for the community to contribute to it and to use it to advance the field of decadal variability. Its success will depend on how much and how well it is used.

New CLIVAR SSG members

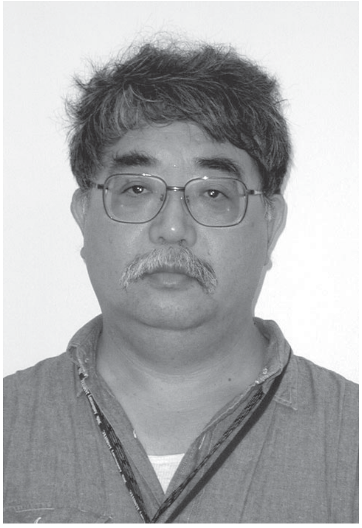
We are pleased to introduce two of the new SSG members: Dr. Ian Simmonds and Dr. Kensuke Takeuchi. Dr. Pedro Silva Dias and Dr. Max Suarez will be introduced in the next issue.

Prof. Dr. Ian Simmonds

Ian received his Ph.D. on 'Modelling the atmosphere with spectral techniques' from Flinders University of South Australia followed by two postdoc affiliations in the Atmospheric Environment Service at Montreal, Quebec, Canada and at the Geophysical Fluid Dynamics Laboratory at Princeton University, Princeton, New Jersey, U.S.A.

Today, Ian is a an Associate Professor in the School of Earth Sciences at The University of Melbourne. He teaches at all levels in the School's Atmospheric and Oceanic Sciences program. He has a very active research program and supervises many students in their Honours and PhD research. Ian has a broad range of research interests in meteorology, climatology, oceanography, climate variability and change, land-atmosphere processes, air-sea interaction, data analysis, and modelling of geophysical systems. In addition to contributions to these fields he has published extensively on the behaviour and influence of Antarctic sea ice, global balances and transport of atmospheric trace gases, circulation of the Antarctic atmosphere, and urban climates.

Prof. Dr. Kensuke Takeuchi



Kensuke is Programme Director of the Frontier Observation Research System for Global Change. His Research interests are on air-sea interaction and climate variations. He was a member of the CLIVAR Upper Ocean Panel from 1999 to 2000. Kensuke got his PhD from the University of Tokyo. Thereafter he moved to the Department of Geophysics at Hokkaido University where he took a Professorship at the Institute of Low Temperature Science.

'Global Change Open Science Conference' - Challenges of a Changing Earth -

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From July 10-13, 2001, the three Global Environmental Change Programmes, i.e., International Geosphere Biosphere Programme (IGBP), International Human Dimensions Programme on Global Environmental Change (IHDP), and the World Climate Research Programme (WCRP), hosted the Global Change Open Science Conference: Challenges of a Changing Earth in Amsterdam, The Netherlands. This major conference was attended by 1700 participants from 94 countries. The scientific programme for the meeting encompassed the range of global change topics including terrestrial and marine productivity, air quality, the global carbon cycle, the global water cycle, global biogeochemistry, land-ocean interactions, climate variability and change, land-use change, biodiversity, the role of technology, and global sustainability. The CLIVAR programme, specifically, and topics of climate variability and predictability were visible throughout the conference in plenary sessions, parallel sessions, poster sessions, and the closing session. In particular there was a plenary on the climate system: prediction, change and variability, a cluster of poster sessions on climate variability and change and parallel sessions devoted to El Niño Southern Oscillation in the context of past and future climate variability, the oceans and climate change, and the atmosphere and global change.

As one might expect from such a broad sweep of topics, the time scales of interest ranged from seasonal to millennial. One example of the intersection of interests in global change was on the decadal time scale and its relation to climate variability, water resources, changes in land use and land cover, societal implications and public policy. As it pertains to decadal variability, this time scale was discussed to be at the upper end of the time scale for political systems (order 4 years), but at the lower end of the time scale of human choice (decades to centuries). Decadal variability and social response is at the nexus of the time scale for present Global Change Programmes (order 15 years), major infrastructural changes (e.g., power, roads, dams, of order 15 years), and the restoration time of regional environmental quality (10-50 years).

In the future, it is anticipated that decadal variability of the coupled climate system will result in ever increasing interaction and feedbacks with social systems as it pertains to water quality/availability and the global carbon cycle. One example of which is the decadal droughts that hit the Nordeste, Brazil. Brazilian literature, going back over 100 hundred years, has frequent references to the drought cycles that hit this region of subsistence farming and a present population of nearly 30 million. During times of particularly severe droughts, there are major emigrations from the region. Those displaced from the Nordeste, often end up in one of two areas, either the major cities and the favelas in São Paulo and Rio de Janeiro, or in Amazonia. Many of the new farmers and loggers involved in the deforestation of the Amazon come from the Nordeste and are merely trying to eke out a living in this frontier. Thus there is an interesting interplay between natural climate variability and social systems such that cyclical droughts induce population migration from one region to another, this results in major changes in land use and land cover, which in the end may come full circle and also feed back to the coupled climate system on the regional, if not, global scale.

Another example, is that of the Bantu tribe in Africa. Lake Naivasha in Kenya also undergoes decadal drought cycles. In the distant past, the Bantu people would migrate to southern Africa when drought hit the region. This was part of their normal adaptation response. However, with the "Balkanization" of countries in the region such population migration has been severely restricted. Their loss of freedom of movement has limited the Bantus' ability to respond to climate variability in the region.

In summary, the interaction between decadal variability and social systems is likely to take on an ever increasing importance in view of the action and response times involved. Climate variability and predictability is seen as a critical basis for a transition towards interdisciplinary sustainability science.

PAGES/CLIVAR met in Amsterdam

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The reconstituted PAGES/CLIVAR Working Group met back to back with the IGBP/WCRP/IHDP Global Change Open Science Conference in Amsterdam, July, 14th, 2001 at the Royal Netherlands Academy of Arts and Sciences (KNAW).

The meeting reviewed PAGES/CLIVAR-related activities during the last year and developed future plans. The following are the main outcomes.

Activities of the past year

- Early last year, a joint PAGES/CLIVAR newsletter was published following the joint workshop on 'Climate of the last Millenium' in Venice, Nov. 1999. This joint venture was a very effective exercise in simulating the PAGES/CLIVAR collaboration.
- A workshop 'Reconstructing Late Holocene Climate' was held April 17-20 2001 in Charlottesville Virginia, organized by M. Mann and H. von Storch. An EOS article was written with a set of research perspectives and suggestions, as well as identifying new areas for international cooperation.
- A session on 'ENSO past and future' at the IGBP/WCRP/IHDP Amsterdam meeting, co-organized by K. Alverson and G. Burgers documented nicely the intersection between the paleo and climate research communities.

Future activities

K. Alverson, director of the PAGES IPO, described future IGBP and PAGES plans in some detail and emphasised the need for PAGES/CLIVAR to provide PAGES with a 5-year plan including goals, timelines and products.

The Working Group reviewed the paleo contribution to the Third Assessment Report of the IPCC and felt that this component should be strengthened in future activities of IPCC.

As a result of a discussion about data issues the panel felt saw a need, in particular for CLIVAR, to strengthen liaison with PAGES to meet the requirements for CLIVAR.

A number of workshops are planned. The PAGES/CLIVAR Working Group will use these activities to foster the collaboration and interaction of the communities.

- An ESF meeting on abrupt climate change will take place in Italy, November 10-15, 2001 organized by J.-C. Duplessy, T. Stocker and K. Alverson.
- J. Overpeck introduced a meeting on paleo-hydrology proposed to be held in 2002 in Tucson, Arizona in 2002. The meeting will focus on the Holocene, primarily the last 2000 years.
- A PAGES/CLIVAR meeting focusing on arctic processes has been proposed by E. Jansen to be held in Norway, September 2002.

In addition, it was pointed out that Open Science Meetings are planned by CLIVAR (late 2003 / early 2004) and PAGES (May 2004 in Bejing).

CLIVAR Calendar

2001/2002	Meeting	Location	Attendance
September 10-14	4th International GEWEX Conference	Paris, France	Open
September 17-20	JSC/CLIVAR Working Group on Coupled Modelling (WGCM), 5th Session	Bracknell, UK	Invitation
September 18-21	Workshop on Advances in the Use of Historical Marine Data: Sea Surface Temperature and Other Key Climate Variables	Boulder, USA	Invitation
October 21-28	IAPSO - IABO 2001: An Ocean Odyssey	Mar del Plata, Argentina	Open
Nov. 4 – 7	First ARTS Open Science Meeting (Annual Records of Tropical Systems)	Noumea, New Caledonia	Open
Nov. 5-7	Working Group on Seasonal to Intterannual Prediction (WGSIP) - 4th Session	Budapest, Hungary	Invitation
Nov. 10-15	Abrupt Climate Change Dynamics	Il Ciocco, Italy	Open
Dec. 10-14	AGU Fall Meeting	San Fransisco	Open
Jan. 13-17	82nd Annual AMS Meeting	Orlando, USA	Open
Feb. 11-15	AGU 2002, Ocean Sciences	Honolulu, USA	Open

Check out our Calendar under: ***http://clivar-search.cms.udel.edu/calendar/default.htm*** *for additional information*

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