3 Chapter three

Understanding climate as a system

Some of the most striking advances in climate change science over the past three to four years have been made by taking a systems perspective

Some of the most striking advances in climate change science over the past three to four years have been made by taking a systems perspective, in which interactions among components of the climate system and feedback processes that highlight potentially important second-order effects have been elucidated. An example is research on the links between climate change and the Hadley Circulation, and the implications of these links for storm tracks, regional precipitation patterns, and modes of natural variability such as the El Niño – Southern Oscillation (Frierson et al. 2007; Lu et al. 2008; Seidel et al. 2008). Much of this new work points in the same direction – that as the 21st century progresses, system-level effects will increasingly amplify rather than dampen the human perturbation of the climate system.

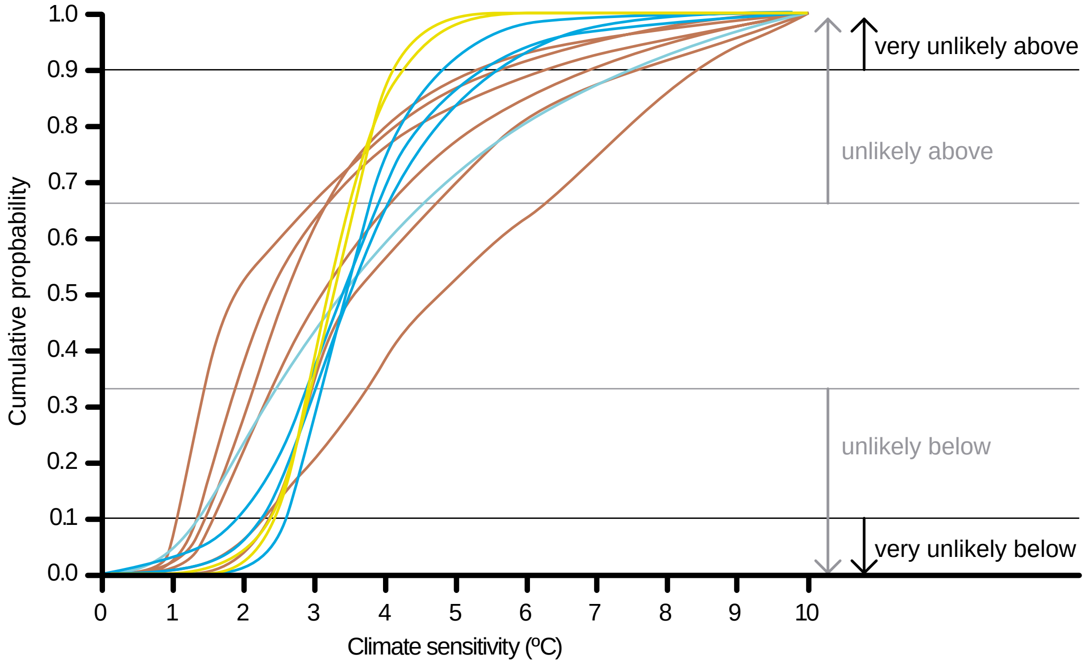
3.1 Climate sensitivity

Climate sensitivity is usually defined as the increase in global mean temperature that would result for a given increase in greenhouse gas forcing once the climate system has reached equilibrium. Climate sensitivity is often defined today as the long-term temperature increase that would result from a doubling of atmospheric CO_2 concentration from pre-industrial, that is, to about 560 parts per million (ppm). The concept is important because, for a given target global mean temperature, climate sensitivity then defines the corresponding atmospheric CO_2 concentration that should become the stabilisation target (Hegerl and Knutti 2008).

Significant progress has been made in determining climate sensitivity, an excellent example of an uncertainty that has been greatly reduced by research. The IPCC's AR4 gives a range of possible sensitivities, with a strong clustering of estimates around the value of 3°C for a doubling of CO_2 (IPCC 2007; Figure 29). This estimate has been obtained from the outputs of numerous simulations by climate models forced with a CO_2 concentration of 560 ppm. The differences among model results arise because of differences within the model formulations in the way in which reinforcing feedbacks within the climate system are handled.

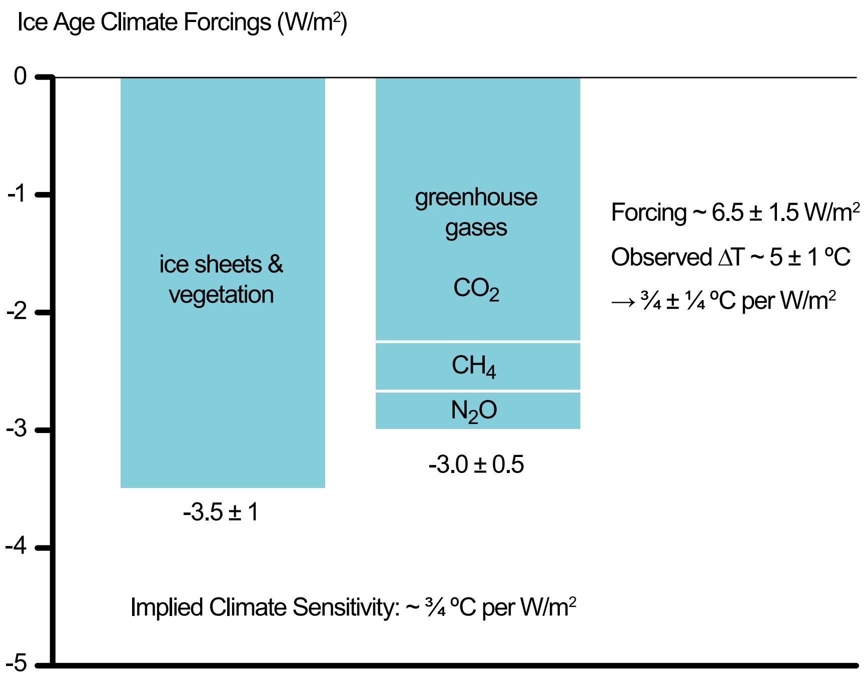
The most important of these reinforcing feedbacks incorporated in all climate models are changes in water vapour concentration and in the amount and types of clouds. Changes in sea-ice extent, which changes the reflectivity of the Earth's surface, are also included. These are so-called "fast feedbacks", in that they operate on timescales comparable to that of the initial radiative forcing. However, the climate system also has "slow feedbacks" that act to warm the climate further given an initial warming; the most important of these are decreased ice area, release of CO_2 from the deep ocean, changed vegetation distribution and inundation of continental shelves and wetlands. These feedbacks, which all change the Earth's radiation balance, may emerge progressively over timescales of centuries and millennia before the climate system reaches equilibrium, but must also be included in the analysis of climate sensitivity. At present none of the climate models include these slow feedbacks.

Figure 29. Climate sensitivity.



Individual cumulative distributions of climate sensitivity from the observed 20th century warming (brown), model climatology (blue) and proxy evidence (light blue). Horizontal lines and arrows mark the edges of the likelihood estimates. (Source: IPCC 2007, including further information on methodology)

Figure 30. Climate forcings.



Climate forcings during ice age 20 ky BP, relative to the present (pre-industrial) interglacial period. (Source: Hansen et al. 2008)

One approach to determining climate sensitivity including both fast and slow feedbacks is to examine how the climate system has responded to changes in CO_2 concentration and ice extent in the past. An analysis of the change in both radiative forcing (Figure 30) and temperature over the 400,000 year Vostok ice core record shows that when ice extent is treated as a forcing rather than a feedback (so that the resulting climate sensitivity includes fast feedbacks only), the sensitivity is well determined at 3°C for a doubling of CO_2. This analysis has substantiated and reduced the uncertainty on the IPCC value. However, when ice extent is treated not as a forcing but as a climate system response, so that both fast and slow feedbacks are included, the climate sensitivity approximately doubles (Hansen et al. 2008). Therefore, the eventual temperature rise in response to a doubling of CO_2 is at least 3°C and likely up to 6°C, depending on the behaviour of the slow feedbacks.

These results are both reassuring and disturbing. They confirm that global climate models, which incorporate only fast feedbacks, have indeed converged on a value for climate sensitivity consistent with the palaeo analysis. Thus, over shorter timeframes of less than a century or two, they are very reliable for projecting changes in global mean temperature. On the other hand, the palaeo-based results imply that the severity of climate change over long timeframes has been significantly underestimated. The slow feedbacks, coupled with the thermal inertia of the ocean, mean that the climate change initially triggered by CO_2 emissions in the 20th and 21st centuries will be irreversible for at least 1,000 years (Solomon et al. 2009). A critical question yet to be answered with confidence is how fast these slow feedback processes can be activated under rapid, sustained radiative forcing (cf. discussion on polar ice sheet dynamics, section 2.1).

The Solomon et al. (2009) analysis cited above emphasises that human-induced climate change will continue for at least 1,000 years following cessation of anthropogenic emissions, with the lower radiative forcing as CO_2 concentration slowly drops largely compensated for by slower heat transfer to the ocean. Thus, if CO_2 levels continue to rise this century, anthropogenic climate change will become effectively irreversible on timescales of relevance for human societies. For example, a peak of 450–600 ppm CO_2 (current CO_2 concentration is about 385 ppm) over the 21st century will likely lead to irreversible rainfall reductions in some regions, as well as to irreversible global sea-level rise that could reach 1.0 m due to thermal expansion alone and could exceed several metres when the contributions from glaciers and ice sheets are included.

3.2 The aerosol masking effect

The direct cooling effect of most aerosols in the atmosphere is becoming better understood, and more detailed estimates of their quantitative effect on radiative forcing are included in the IPCC AR4 (2007). Their overall cooling effect is roughly equivalent to the warming effect of the non-CO_2 greenhouse gases. However, significant uncertainties still surround the indirect effects of aerosols via clouds and precipitation and thus their overall effect on climate. This uncertainty is important, as efforts to reduce local and regional air pollution (the major sources of the aerosol load in the atmosphere) could lead to a surge in warming without any additional increase in greenhouse gases (Andreae et al. 2005). In effect, the aerosol masking effect increases the level of “committed climate change” that is in the pipeline with current levels of greenhouse gases.

A new analysis has attempted to pin down more precisely the level of committed climate change that is implied by the aerosol masking effect (Ramanathan and Feng 2008). Using the IPCC AR4 (2007) estimates of greenhouse gas forcing and climate sensitivity, and revised estimates of the aerosol cooling effect (Ramanathan and Carmichael 2008), the analysis argues that even if greenhouse gas concentrations could be fixed at their 2005 levels, the additional warming caused by cleaning up aerosol pollution and reaching atmosphere–ocean equilibrium would be 1.6°C. This represents the best estimate of “committed temperature increase”. This would yield an overall temperature rise compared to pre-industrial levels of 2.4°C (with a range of 1.4 to 4.3°C), enough to eliminate Arctic sea ice, reduce or eliminate the Himalayan–Tibetan glaciers and possibly push the Greenland ice sheet past the point that will trigger irreversible loss of the ice sheet. These conclusions imply that, to avoid these risks, society will need to achieve the most vigorous of the IPCC emissions scenarios (peak/decline/negative emissions late in the 21st century).

3.3 Carbon cycle feedbacks

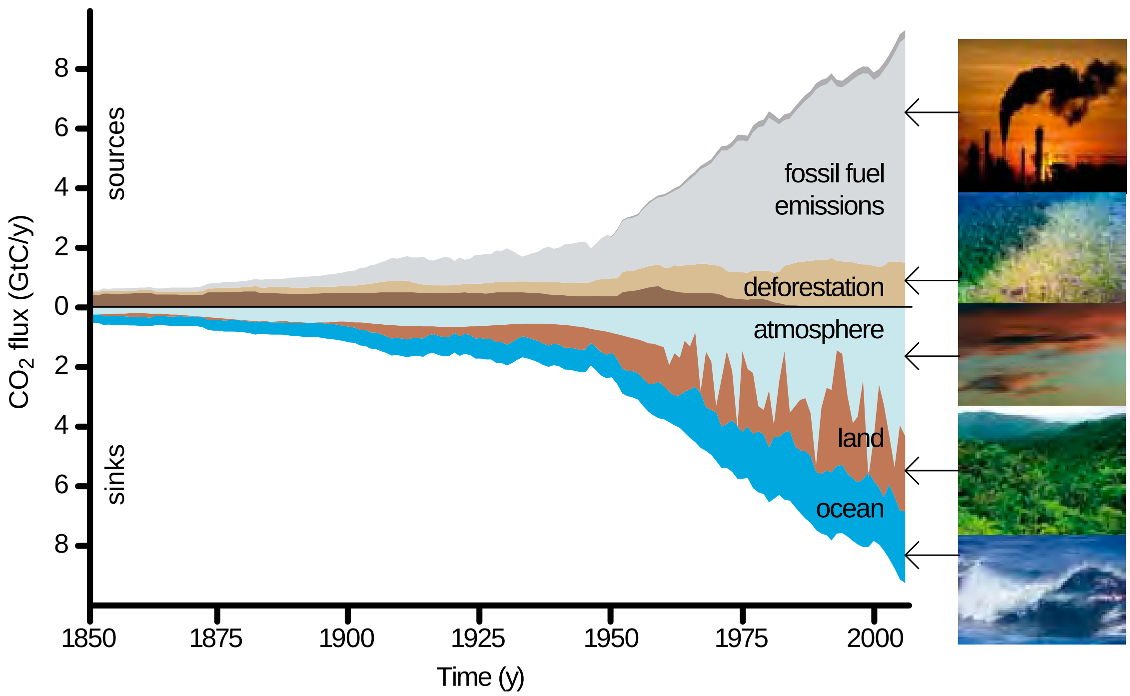
Less than half of the CO_2 emitted to the atmosphere by human activities (including fossil fuel combustion and land-use change) remains there; the rest is taken up by natural sinks on land and in the oceans (Figure 31). A model intercomparison (Friedlingstein et al. 2006) showed the potential for these sinks to weaken as a result of a warming climate, allowing further CO_2 to remain in the atmosphere and thus acting as a reinforcing (positive) feedback to climate change.

Observations over five decades of the fraction of CO_2 emissions taken up by land and ocean sinks (Canadell et al. 2007; Figure 32) show very high interannual variability in the uptake fraction by the land sink but no long-term trend. The fraction taken up by the oceanic sink, on the other hand, has shown a significant downward trend over the past 40 years, from absorbing about 32% of anthropogenic emissions in 1960 to about 26% now. This weakening is due to several factors, including warming and acidification in the upper ocean. An additional possible contributing process is a strengthening of the circumpolar winds across the Southern Ocean, leading to enhanced ventilation of carbon-rich deep waters and consequent slower uptake of atmospheric CO_2 (Le Quéré et al. 2007), although others question this (Matear and Lenton 2008; Law et al. 2007; Böning et al. 2008). The overall result is that the combined natural (ocean plus land) CO_2 sink is decreasing as a proportion of CO_2 emissions, and therefore its buffering effect on the rate of climate change is reducing.

In addition to the weakening in the efficiency of existing natural sinks of carbon, there is concern that some new natural sources of carbon could become activated as the climate warms further. Much of the concern is focused on the northern high latitudes, and on the carbon stored in frozen soils (permafrost) (Figure 33). Recent observations of an upturn in methane (CH_4) emissions (Figure 34) add to this concern.

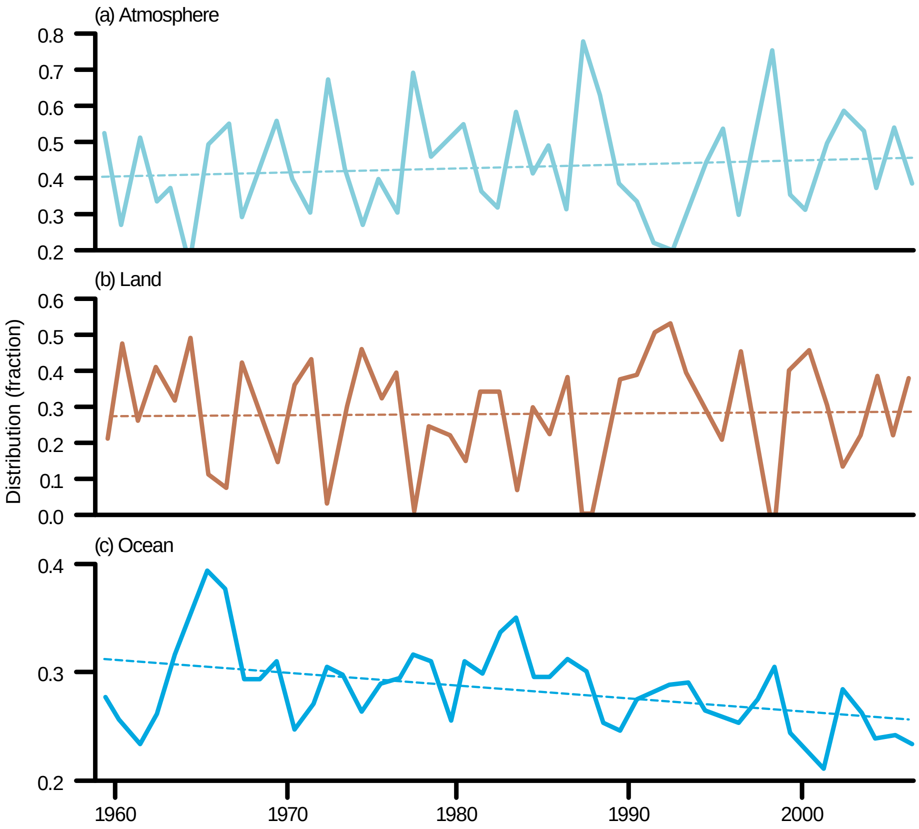
An updated estimate of the amount of carbon stored in permafrost in the northern high latitudes has highlighted the global importance of this potential source (Tarnocai et al. 2009). The new estimate is 1,672 billion tonnes, about double the value reported in previous analyses. This carbon alone would account for about half of the estimated global belowground organic carbon pool, and is about twice as much as the total amount of carbon in the atmosphere. Schurr et al. (2008) examined the processes by which this permafrost carbon

Figure 31. Changes in the global carbon budget from 1850 to 2000.



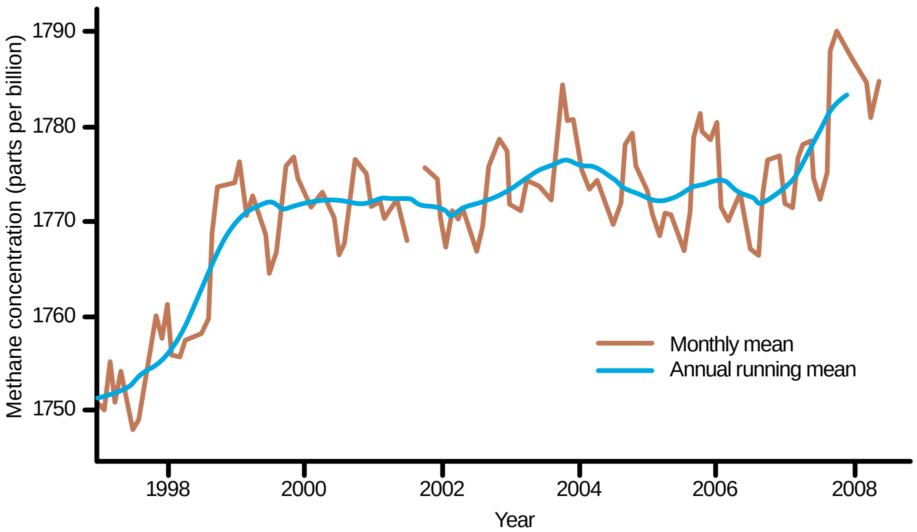
CO_2 emissions to the atmosphere (sources) as the sum of fossil fuel combustion, land-use change, and other emissions (upper); the fate of the emitted CO_2, including the increase in atmospheric CO_2 pus the sinks of CO_2 on land and in the oceans (lower). Flux is in Pg y^{-1} carbon. (Source: Adapted from Canadell et al. 2007, with additional data from C. Le Quéré and the Global Carbon Project)

Figure 32. Fraction of the total emissions that remains in the (A) atmosphere, (B) the land biosphere, and (C) the ocean.



(Source: Canadell et al. 2007)

Figure 34. The average atmospheric concentration of methane from 1997 to 2008.



The graph demonstrates the sharp increase in 2007. (Source: From M. Rigby, CSIRO, based on data from the Advanced Global Atmospheric Gases Experiment)

Figure 33. Latitudinal zonation of permafrost.

(Source: Brown et al. 1998)

could be transferred to the atmosphere. Thawing of permafrost can occur both gradually and abruptly, and carbon can be lost from thawing permafrost to the atmosphere as either CO_2 or CH_4, depending on whether it is decomposed under aerobic or anaerobic conditions. As climate warms, there are also some processes associated with ecosystem change in the high latitudes of the northern hemisphere that can lead to uptake of carbon from the atmosphere (e.g. longer growing seasons), potentially counteracting loss of carbon from permafrost soil; however, these processes appear to be incapable of forming a sink of sufficient magnitude in the short timeframes over which permafrost carbon could be lost. The net effect is likely to be a positive feedback to a warming climate.

Although the impacts of a warming climate on natural land and ocean carbon sinks are complex, with both positive and negative feedbacks possible, the observational evidence so far and model projections suggest that the positive (reinforcing) feedbacks will dominate. If such outcomes eventuate, reduction of anthropogenic emissions will have to be even deeper to achieve stabilisation of atmospheric CO_2 at a given level.

3.4 Understanding the climate of the past

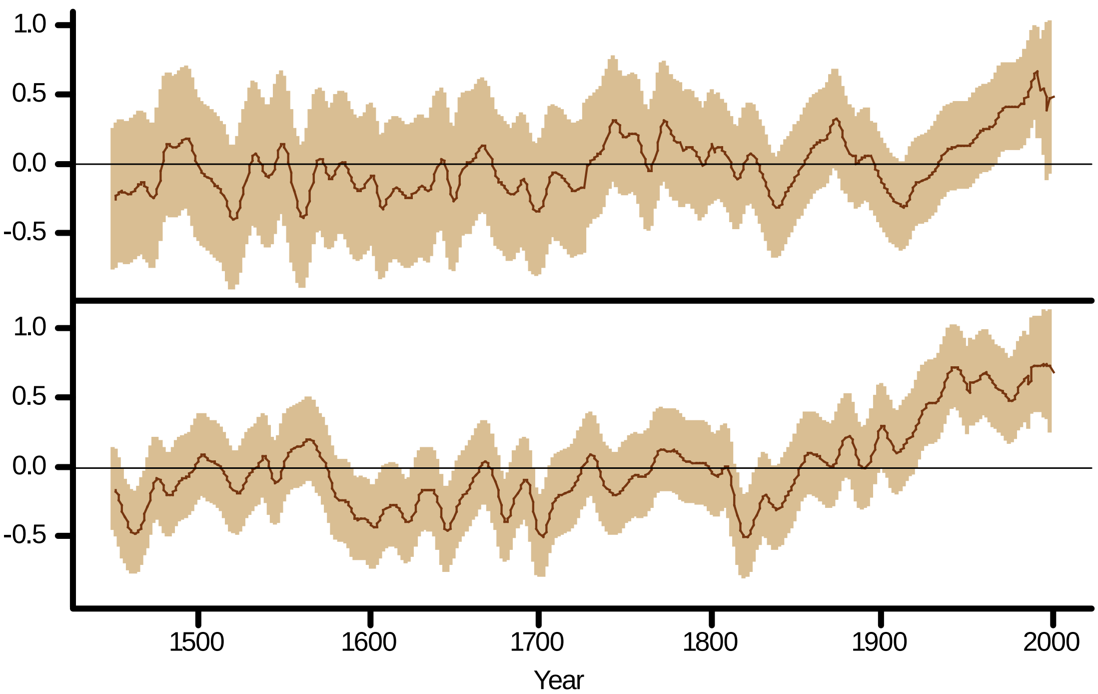
Palaeo-climatic studies are rapidly increasing in importance as tools to place contemporary climate change into a longer-term context (Otto-Bleisner et al. 2009). Understanding the climate of the past yields information on patterns of natural variability, gives insights into critical processes in the climate systems, and provides data for testing climate models in situations when radiative forcing and climatic parameters were significantly different from today. Palaeo-climatic research played a stronger role in the IPCC AR4 (2007), but much of the information was limited to the northern hemisphere. Despite recent progress, much remains to be done in the Australian region, and in the southern hemisphere more broadly, to attain the same level of understanding of past climatic changes as in the north.

One of the most obvious challenges facing the southern hemisphere research community is to build a millennial scale temperature reconstruction similar to that of the northern hemisphere (Mann et al. 1999; Mann and Jones 2003), the so-called “hockey stick” temperature reconstruction. Some progress has been made towards that goal, with the development of a 550-year temperature reconstruction for the southern hemisphere based on 14 proxy records (tree ring widths, coral calcification, isotope records of carbon and oxygen in ice cores) across South America, Africa, Australia and New Zealand (Turney and Duncan 2008). The results (Figure 35) showed a complex pattern of natural variability different from that of the northern hemisphere, but both hemispheres showed the same significant warming trend through the 20th century. In both cases, the 20th century warming was unusual compared to the longer-term pattern of natural variability shown in the reconstructions.

As noted in section 2.2, changes in water availability comprise one of the most important consequences of climate change for Australia. Palaeo-climatic research is now providing more insights into the ways in which important modes of natural variability have behaved over long time periods, and how they may be affected now by anthropogenic climate change.

A detailed record of the behaviour of the El Niño – Southern Oscillation phenomenon (ENSO) from AD 1525 to 1982 has now been reconstructed (Braganza et al. 2009), building on a large set of

Figure 35. Southern hemisphere (top) and northern hemisphere (bottom) temperature reconstructions.

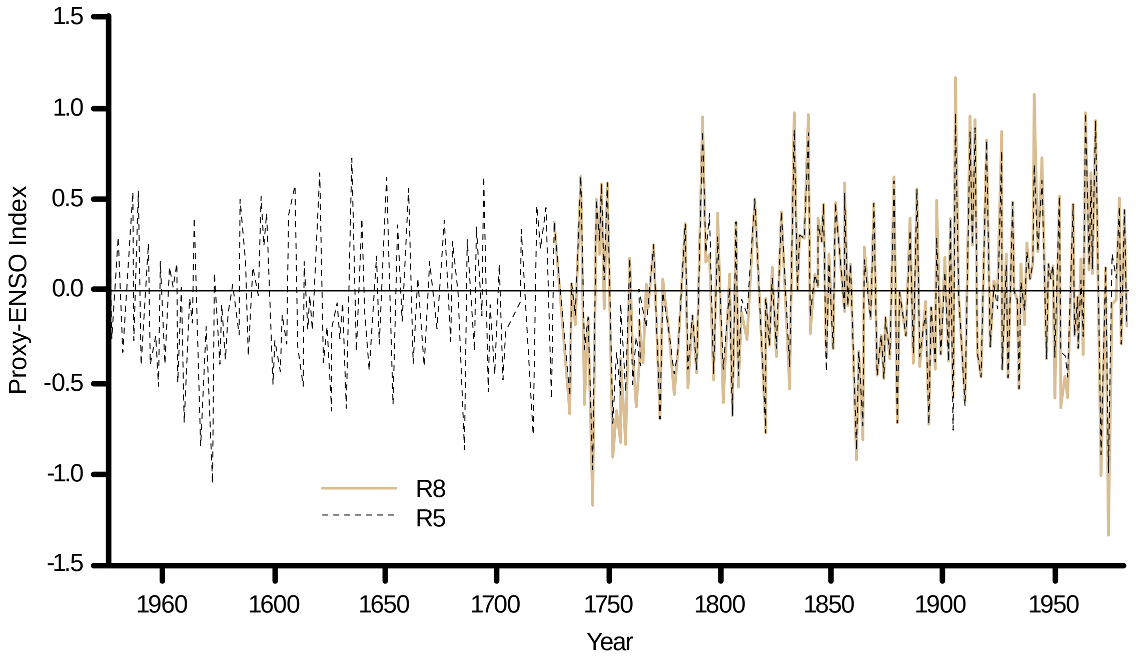


The dark lines are the mean reconstructions and the brown shading is approximate 95% confidence intervals around the means. (Source top graph: Turney and Duncan 2008, source bottom graph: Osborn and Briffa 2006) climate proxies from around the entire Pacific Basin. The results (Figure 36) show that its variability was relatively low during the 16th, 17th and early 18th centuries, but that high-frequency variability (ca 2–4 year intervals) has increased over the last 200 years compared with the earlier period. Perhaps somewhat surprisingly, there is no pronounced change in ENSO variability in the 20th century – and hence no apparent change in ENSO behaviour as a result of anthropogenic climate change. However, the reconstruction stops at 1982, and thus does not include the recent period of rapid warming. Further research is required to determine if ENSO behaviour has changed in the late 20th and early 21st centuries.

Given the growing recognition of the importance of the Indian Ocean Dipole (IOD) phenomenon in influencing rainfall in south-east Australia, a better understanding of its behaviour in the past is essential. A study based on coral geochemical records in the equatorial eastern Indian Ocean over the past 6,500 years shows enhanced cooling and drying in the region early in the record – the mid Holocene – compared to the present (Abram et al. 2007). The work has uncovered a strong link between the Indian Ocean Dipole and the strength of the Asian monsoon, in which a stronger monsoon is coupled with drier conditions in the eastern Indian Ocean (Overpeck and Cole 2007). Thus, a strengthening of the Asian monsoon, which occurred during the mid Holocene and is expected with a warming climate (Ashrit et al. 2001; Hu et al. 2000), would lead to drier conditions in western Indonesia and eastern Australia. A further coral-based study of the behaviour of the Indian Ocean Dipole since 1846 (Abram et al. 2008) shows an increase in the frequency and strength of Indian Ocean Dipole events (Figure 37) in the 20th century, which may be linked to anthropogenic climate change. Furthermore, the trend in Indian Ocean Dipole behaviour is associated with the development of a direct link to the Asian monsoon and a weakening of the historical relationship between the El Niño – Southern Oscillation and the Indian Ocean Dipole.

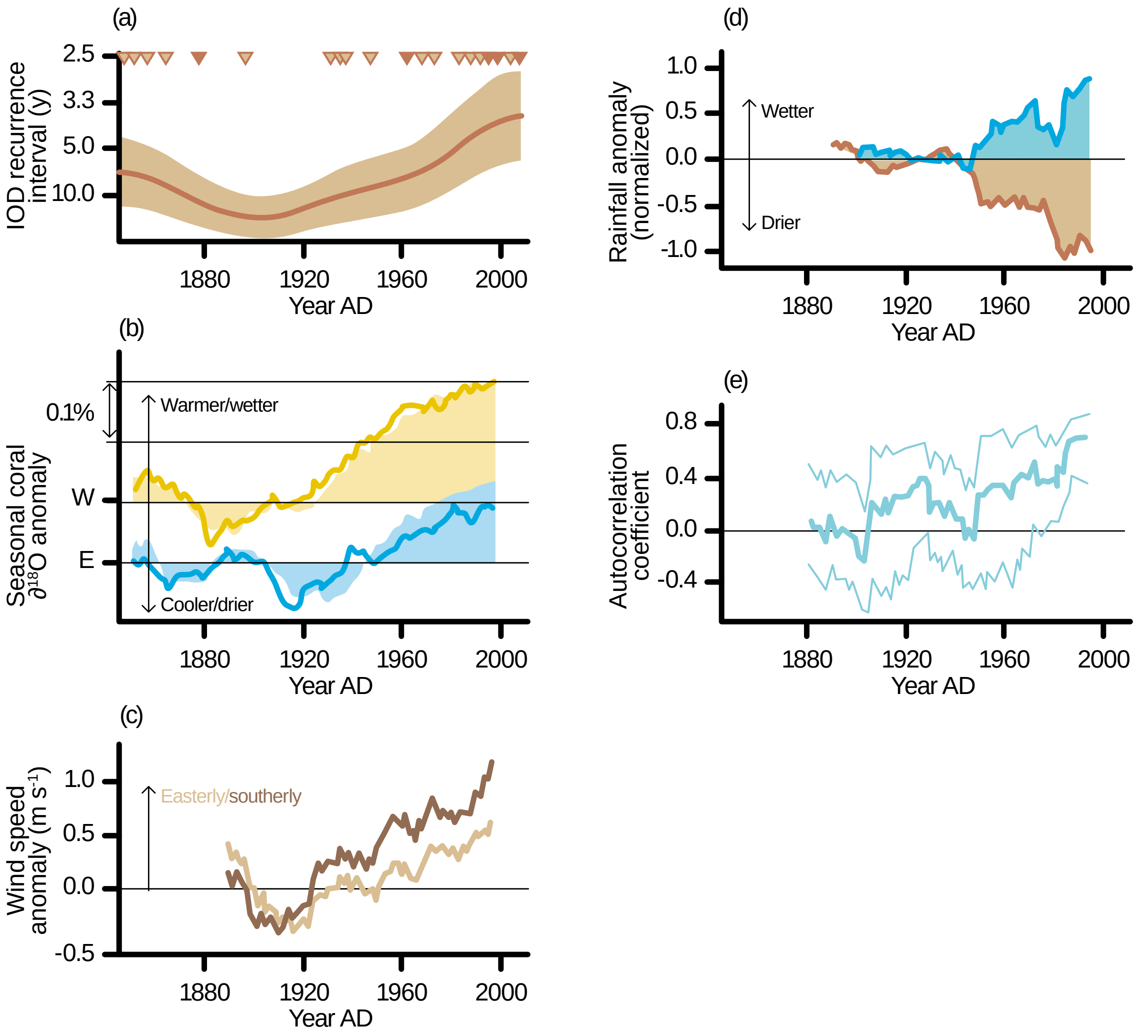
In general, this palaeo-research supports the conclusion, described in section 2.2, that the drought in south-east Australia is probably linked, at least partly, to climate change, but it is much more difficult to establish a link between climate change and the drying trend further north along Australia's east coast.

Figure 36. Past behaviour of the El Niño Southern Oscillation (ENSO).



Proxy ENSO indices R5 (dotted) and R8 (brown) for the period 1525–1982. (Source: Braganza et al. 2009)

Figure 37. 20th-century intensification of Indian Ocean Dipole (IOD).



(a) Occurrence rate of IOD events (brown curve); the triangles show moderate (light) and strong (dark) IOD events. (b) 25-year moving averages of eastern (blue) and western (yellow) coral delta-18 O during the IOD season (July–November, lines) and non-IOD season (March–May, shading) relative to pre-1940 means. (c) 25-year moving averages of IOD-season zonal (dark, inverted) and meridional (light) wind speed over the eastern IOD upwelling region, relative to pre-1940 means. (d) 25-year moving averages of IOD-season rainfall in western Indonesia (brown) and eastern Africa (blue), normalised relative to the common 1910–1940 interval. (e) 25-year moving correlation between ENSO-dependent residuals of the IOD and Asian monsoon (solid curve) with a 95% Monte Carlo confidence window (light blue). (Source: Abram et al. 2008)

4 Chapter four

Over-the-horizon research

The frontiers of climate research are rapidly spreading into what is often called Earth System science. Although definitions of Earth System science vary, the concept is usually associated with the interactions of climate with other features of the planetary environment that interact with climate, with a long-term perspective into both the past and the future, and – importantly – with the inclusion of human and societal dynamics as a fully integrated, interactive part of the Earth System.

4.1 Seamless prediction from weather to climate

Weather forecasting has improved markedly over the past decade due to a number of factors, including better understanding of the weather/climate system and the capability to assimilate large amounts of observational data into the modelling framework in near real-time. Also within the last decade or so the capability to predict interannual variability in climate has been developed with the advent of El Niño Southern Oscillation forecasting. Global Climate Models (GCMs) can predict the state of the climate system a century into the future, given external forcing factors. These developments provide a platform on which to build a seamless weather/climate prediction system that can generate forecasts from hours to decades to a century or two.

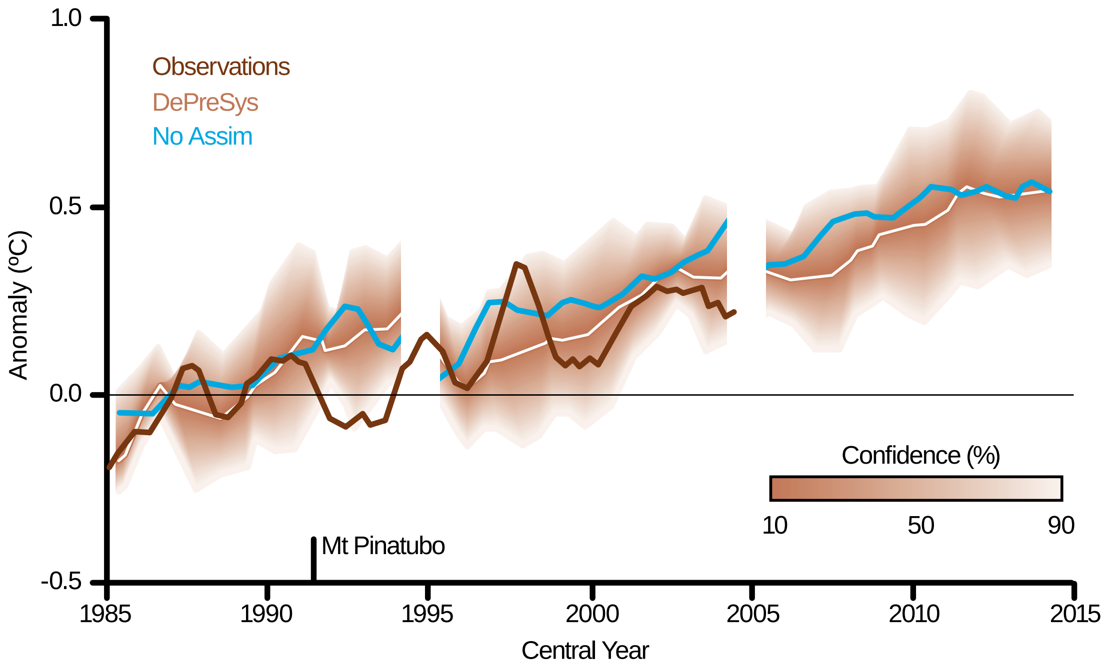
The missing scale in current modelling expertise is from a few years to a few decades, and there is rapidly growing interest in how feasible predictions on this scale actually are and how best to achieve an acceptable level of predictability. This intermediate scale of prediction is particularly important for adaptation to climate change. Many of the changes in policy and management that will need to be made in a wide range of sectors operate on this timescale.

Weather forecasting has improved markedly over the past decade…

The first attempts at tackling this intermediate scale of prediction have been made by decadal-scale GCM simulations that incorporate better understanding of important modes of interannual variability, anomalies in ocean heat content and fluctuations in thermohaline circulation. One of these modelling systems, based on the Hadley Centre GCM, has been used recently to predict the behaviour of the climate system for the next decade (Smith et al. 2007). One of the key features of the system is that it takes into account the observed state of the atmosphere and ocean (i.e. initial condition information), which is necessary to predict internal variability of the climate system, as well as projected changes in natural (solar irradiance) and anthropogenic (greenhouse gas emissions) drivers of climate.

The modelling system was tested in hindcast mode from 1985 and then into the future to 2015 (Figure 38; Smith et al. 2007). The model has been able to predict the levelling out of atmospheric warming from about 2002 to 2009 but then predicts the warming to resume around 2010, or 2011 with half of the years after 2009 predicted to be warmer than 1998, the warmest year on record so far. These simulations demonstrate the importance of ocean circulation in modulating climate over decadal timeframes and the importance of the observed ocean state for the initial conditions for these simulations. The oceans absorb more than 80% of the additional energy at the Earth's surface due to rising atmospheric concentrations of greenhouse gases (IPCC 2007).

Figure 38. Seamless climate prediction.



Globally averaged annual mean surface temperature anomaly (relative to 1979–2001) forecast by a decadal climate prediction system (DePreSys) starting from June 2005. The confidence interval (brown shading) is diagnosed from the standard deviation of the DePreSys ensemble, assuming a t distribution centred on the ensemble mean (white curve). Also shown are hindcasts from DePreSys and an ensemble mean from hindcasts by an identical modelling system that did not assimilate the observed state of the atmosphere or ocean (blue curves). Observations from the Hadley Centre data set are in brown. (Source: Smith et al. 2007)

4.2. Tipping elements in the climate system

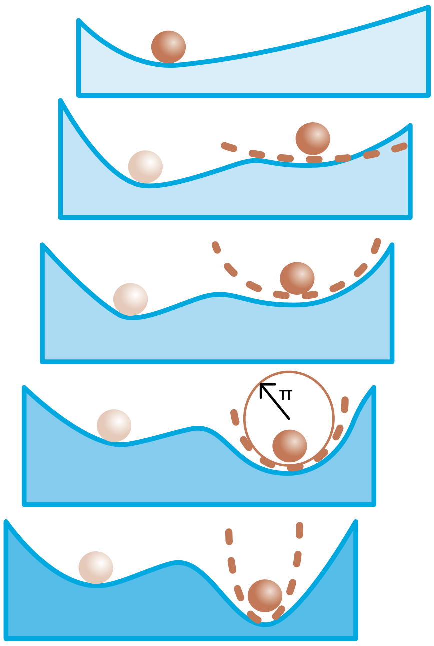
One of the most dangerous features of the climate system in terms of impacts on societies is the potential for abrupt and/or (essentially) irreversible changes when thresholds are crossed. Threshold/abrupt change behaviour occurs when a small perturbation to a control variable can cause a rapid and unexpectedly large change in a system, altering its state or direction of development (Figure 39; Lenton et al. 2008). Perhaps the best-known example of such behaviour in the climate system is the potential abrupt shutdown of the North Atlantic Thermohaline Circulation with an influx of freshwater on the surface of the regional ocean (Clark et al. 2002).

Although the existence of thresholds and abrupt changes in the climate system has been known for some time, research over the past couple of years has added significantly to the knowledge base. A recent analysis of “tipping elements” in the climate system has identified a set of large-scale components of the climate system that could undergo abrupt or irreversible change under anthropogenic forcing (Lenton et al. 2008; Figure 40; Table 1). One of the criteria for inclusion in the list is that the tipping element has significant consequences for human well-being should it be altered. In addition, the tipping element should be capable of being triggered this century and undergo a qualitative change this millennium.

An example of such a tipping element is the Greenland ice sheet. It is possible that the threshold leading to its eventual disappearance could be crossed later this century (Gregory and Huybrechts 2006), leading to a sea-level rise of around 7 m that would likely be realised within this millennium. It is also possible that the West Antarctic Ice Sheet could cross a threshold this century, but at a higher level of mean global temperature than for Greenland (Lenton et al. 2008). Such changes would be irreversible in any timeframe meaningful for human societies.

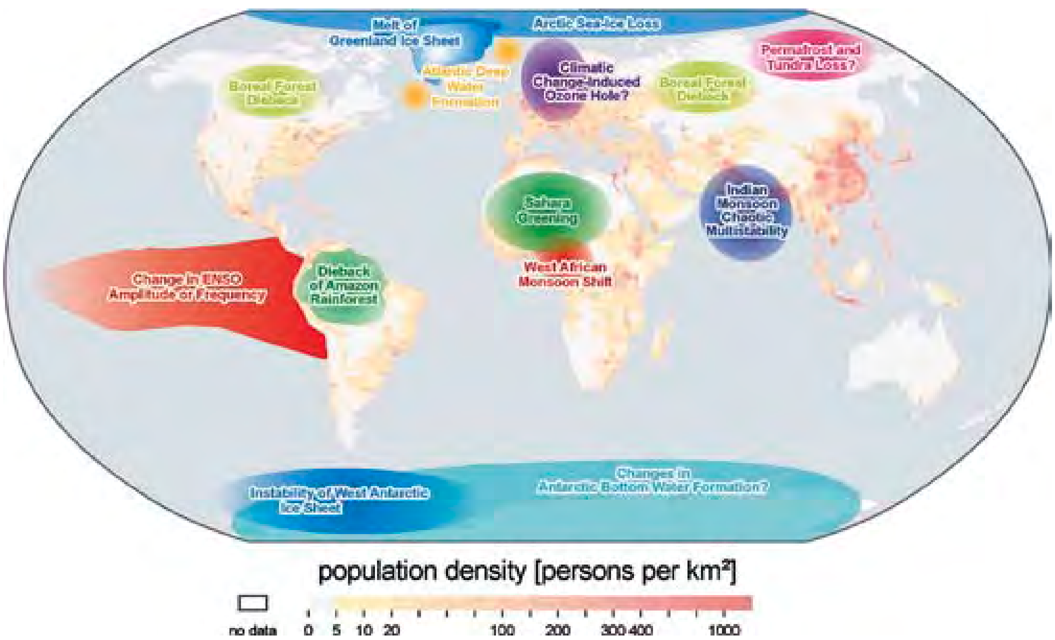
Another example is the Indian monsoon system, on which more than 1 billion people depend for their food production and, ultimately, their water supplies. This system can undergo a transition very rapidly, in only one year, leading to drought and a significant reduction in the number of humans that can be supported in the region (Lenton et al. 2008, Table 1; Zickfeld et al. 2005). Such an abrupt change would trigger a human catastrophe of disturbing proportions.

Figure 39. Schematic illustrating poximity to a tipping point.



The potential wells represent stable attractors, and the ball, the state of the system. Under gradual anthropogenic forcing (progressing from bottom left to top right), the right potential well becomes shallower and finally vanishes (threshold), causing the ball to abruptly roll to the left. (Source: Lenton et al. 2008)

Figure 40. Map of potential policy-relevant tipping elements in the climate system, overlain on global population density.



Elements shown could exhibit threshold-type behaviour in response to anthropogenic climate forcing, where a small perturbation at a critical point qualitatively alters the future fate of the system. They could be triggered this century and would undergo a qualitative change within this millennium. Question marks indicate systems whose status as tipping elements is particularly uncertain. (Source: Lenton et al. 2008)

Table 1: Policy-relevant potential tipping elements in the climate system.

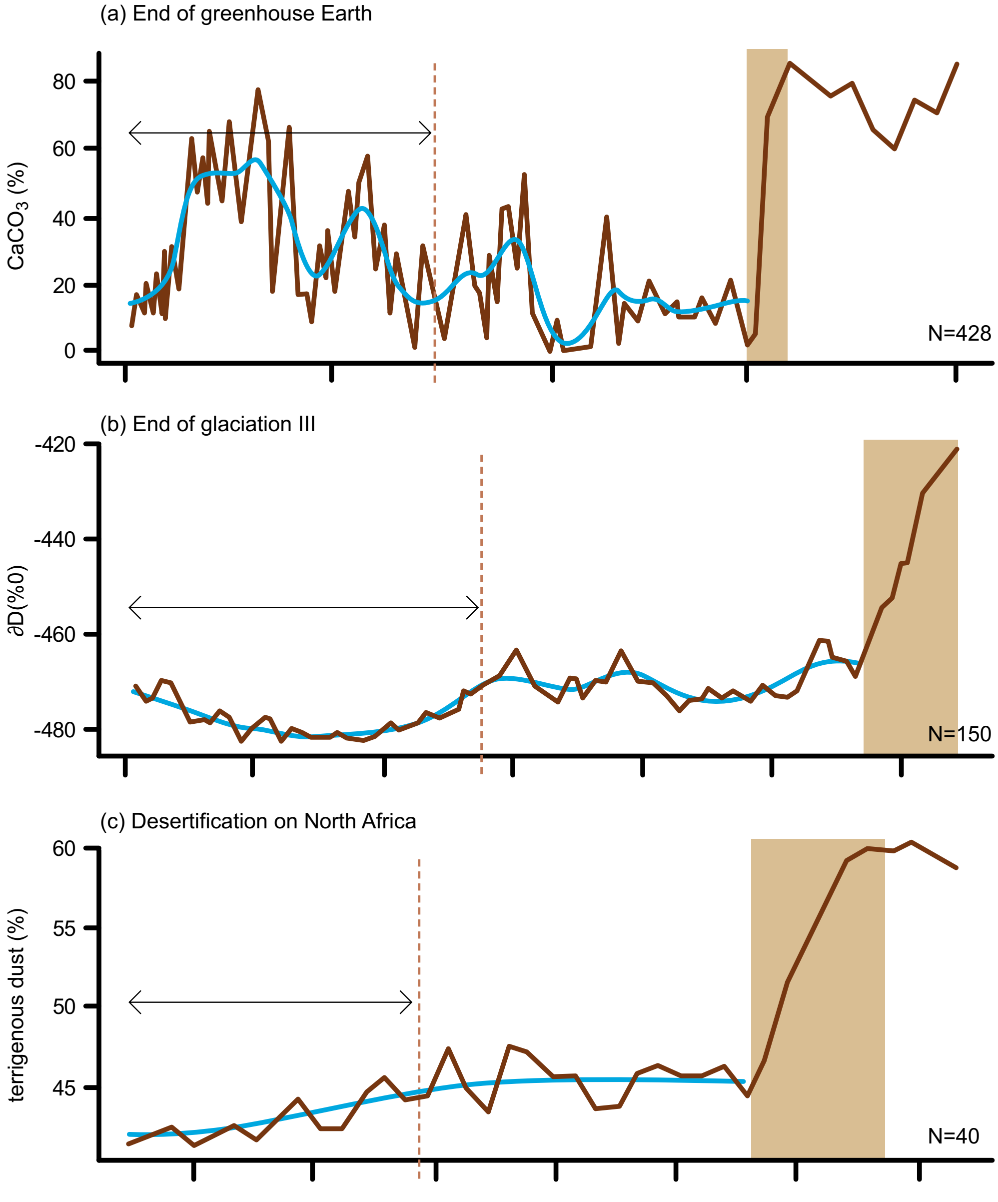
Tipping element	Feature of system, *F* (direction of change)	Control parameter(s), ρ	Critical value(s),† ρ_{crit}	Global warming†,‡	Transition timescale,† *T*	Key impacts
Arctic summer sea-ice	Areal extent (−)	Local ΔT_{air}, ocean heat transport	Unidentified§	+0.5–2°C	≈10 yr (rapid)	Amplified warming, ecosystem change
Greenland ice sheet (GIS)	Ice volume (−)	Local ΔT_{air}	+≈3°C	+1–2°C	>300 yr (slow)	Sea level +2–7 m
West Antarctic ice sheet (WAIS)	Ice volume (−)	Local ΔT_{air}, or less ΔT_{ocean}	+≈5–8°C	+3–5°C	>300 yr (slow)	Sea level +5 m
Atlantic thermohaline circulation (THC)	Overturning (−)	Freshwater input to N Atlantic	+0.1–0.5 Sv	+3–5°C	≈100 yr (gradual)	Regional cooling, sea level, ITCZ shift
El Niño–Southern Oscillation (ENSO)	Amplitude (+)	Thermocline depth, sharpness in EEP	Unidentified§	+3–6°C	≈100 yr (gradual)	Drought in SE Asia and elsewhere
Indian summer monsoon (ISM)	Rainfall (−)	Planetary albedo over India	0.5	N/A	≈1 yr (rapid)	Drought, decreased carrying capacity
Sahara/Sahel and West African monsoon (WAM)	Vegetation fraction (+)	Precipitation	100 mm/yr	+3–5°C	≈10 yr (rapid)	Increased carrying capacity
Amazon rainforest	Tree fraction (−)	Precipitation, dry season length	1,100 mm/yr	+3–4°C	≈50 yr (gradual)	Biodiversity loss, decreased rainfall
Boreal forest	Tree fraction (−)	Local ΔT_{air}	+≈7°C	+3–5°C	≈50 yr (gradual)	Biome switch
Antarctic Bottom Water (AABW)*	Formation (−)	Precipitation–Evaporation	+100 mm/yr	Unclear¶	≈100 yr (gradual)	Ocean circulation, carbon storage
Tundra*	Tree fraction (+)	Growing degree days above zero	Missing‖	—	≈100 yr (gradual)	Amplified warming, biome switch
Permafrost*	Volume (−)	$\Delta T_{permafrost}$	Missing‖	—	<100 yr (gradual)	CH_4 and CO_2 release
Marine methane hydrates*	Hydrate volume (−)	$\Delta T_{sediment}$	Unidentified§	Unclear¶	10^3 to 10^5 yr (>T_E)	Amplified global warming
Ocean anoxia*	Ocean anoxia (+)	Phosphorus input to ocean	+≈20%	Unclear¶	≈10^4 yr (>T_E)	Marine mass extinction
Arctic ozone*	Column depth (−)	Polar stratospheric cloud formation	195 K	Unclear¶	<1 yr (rapid)	Increased UV at surface

N, North; ITCZ, Inter-tropical Convergence Zone; EEP, East Equatorial Pacific; SE, Southeast.
* See *SI Appendix 2* for more details about the tipping elements that failed to make the short list.
† Numbers given are preliminary and derive from assessments by the experts at the workshop, aggregation of their opinions at the workshop, and review of the literature.
‡ Global mean temperature change above present (1980–1999) that corresponds to critical value of control, where this can be meaningfully related to global temperature.
§ Meaning theory, model results, or paleo-data suggest the existence of a critical threshold but a numerical value is lacking in the literature.
¶ Meaning either a corresponding global warming range is not established or global warming is not the only or the dominant forcing.
‖ Meaning no subcontinental scale critical threshold could be identified, even though a local geographical threshold may exist.

(Source: Lenton et al. 2008)

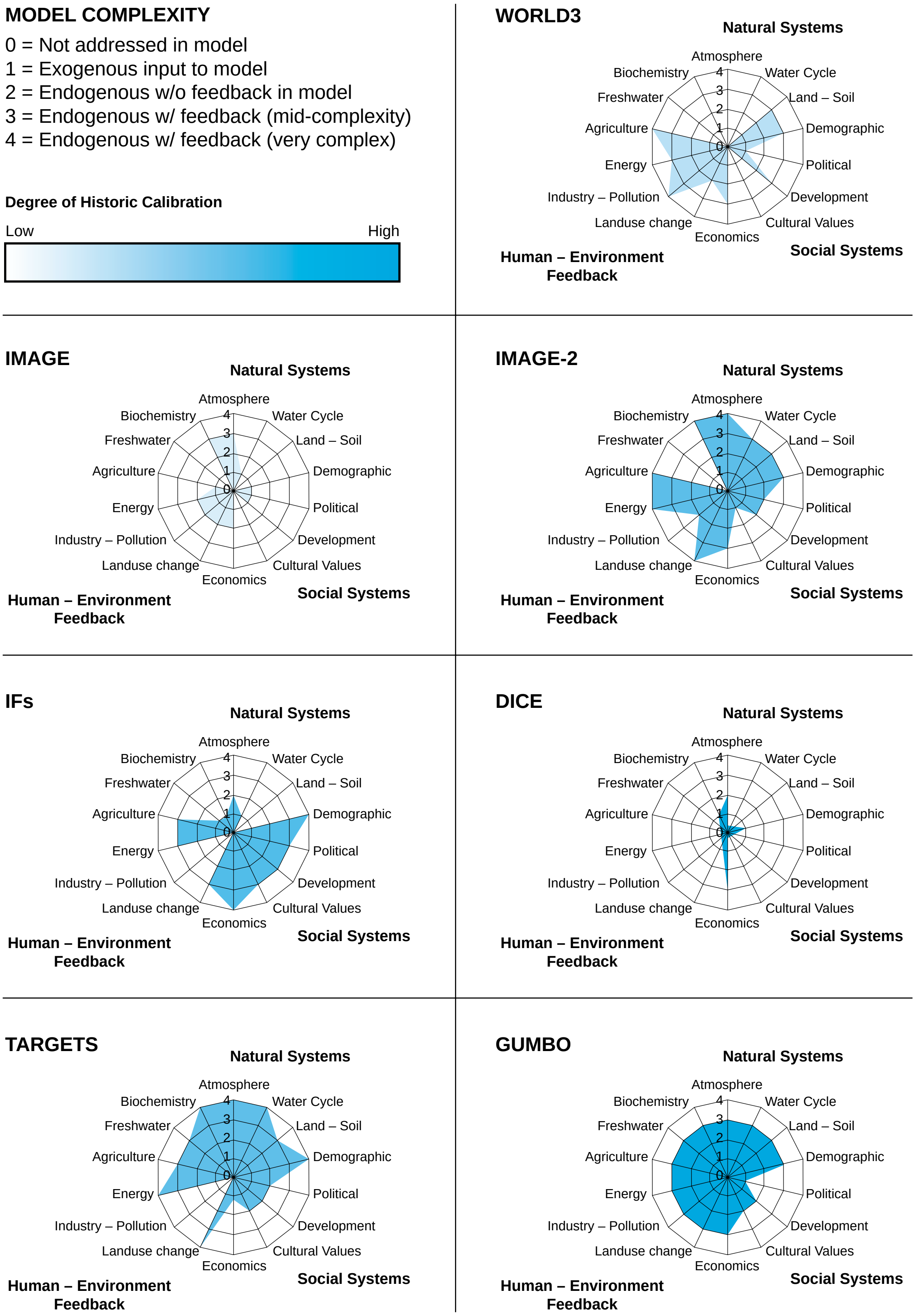
An obvious question arises: can we anticipate the approach to such a threshold early enough so that crossing it and triggering the abrupt change can be avoided? Some intriguing research examining abrupt changes in the past suggests that we can indeed anticipate when an abrupt change is nearing (Dakos et al. 2008). Analysis of several abrupt changes of the past shows that there is always a slowing down of fluctuations (natural variability) in the system prior to the threshold being crossed (Figure 41). In each case there was increased autocorrelation in system-level fluctuations as the threshold was approached, a characteristic that appears to be a universal warning signal for the imminent triggering of abrupt change. If this hypothesis holds true for climate, it would be an exceptionally powerful tool for humanity to anticipate and avoid the nasty surprises that are likely inherent in the behaviour of the climate system. The crucial question remains as to whether the warning signs can be discerned early enough (e.g. on timeframe of decades or centuries) to allow policy and management the opportunity to reduce the forcing before the threshold is crossed.

Figure 41. Abrupt climate shifts.



Three reconstructed time series of abrupt climate shifts in the past (A) the end of the greenhouse Earth; (B) the end of glaciation III, and (C) the desertification of North Africa. In each case the dynamics of the system slow down before the transition. The grey areas identify transition phases. (Source: Dakos et al. 2008, including sources of data and details of methodology)

Figure 42. Complexity of Integrated Global Models (IGMs).



Amoeba diagram of complexity with which IGMs capture socio-economic systems, natural systems, and feedbacks. (Source: Costanza et al. 2007)

4.3 Putting humans into Earth System models

One of the most challenging research tasks ahead is to couple economic and social dynamics with the biophysical climate system in an interactive way. At present human actions are usually represented as an outside force perturbing the "natural" climate system via a greenhouse gas emissions scenario, or climate is simply represented by a damage function related to temperature embedded in a much more complex economic model. Achieving a balance between the human and biophysical components of a global-scale model has proven to be difficult.

Perhaps the most successful coupled models to date are Integrated Assessment Models, which are normally used to explore the consequences of various climate change scenarios for society. They are often organised around a number of important sectors – energy, industry, agriculture, trade, for example – and use a simplified version of a GCM for the climate model. A broader genre of models that attempt to simulate the interactive relationship between humans and the environment are often called Integrated Global Models. A recent review of such models (Costanza et al. 2007; Figure 42) again demonstrates the challenge to build models that are not strongly skewed towards either socio-economic systems or the biophysical climate system. Note in Figure 42 that climate is represented simply by "atmosphere" in many of these models.

In terms of human dynamics, a challenge for future modelling efforts is to capture the complexity of the ways in which societies are responding to climate change and will do so in the future. New approaches aimed at meeting this challenge include massive agent-based modelling, social network theory, game theory, evolutionary psychology and complex systems theory, or some combination of these.

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