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TEMPERATURE LIMITS IN THE CLIMATE SYSTEM

REGINALD E NEWELL AND ZHONGXIANG WU

Department of Earth, Atmospheric and Planetary Sciences
Room 54-1824, *Massachusetts Institute of Technology, Cambridge, MA* 02139, *USA*

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Physical processes in the climate system appear to limit maximum and minimum temperatures in various regions of the ocean and atmosphere. Possible limits in four regions are discussed here: the surface ocean in the tropical west and east Pacific, the tropical free troposphere, and the lower tropical stratosphere. The processes are deduced from recent data sets but show some consistency with data from the past ice age. The lower limit on surface temperatures at high latitudes during ice age conditions is also examined.

Key Words: Climate; Limiting Temperatures; Warm Pool; Ages; Volcanic Activity; Associated Temperature Change

Introduction

Maps of tropical Pacific surface temperature (Fig. 1) show a region in the tropical west Pacific where temperatures are close to 29°C throughout the year, following the sun at lag between the hemispheres. Why is this value 29°C and not 25°C or 35°C? Why does the tropical east Pacific also never exceed this value? Likewise the tropical east Pacific rarely goes below 18°C. What controls this limit?

During an El Niño the extra energy flux into the atmosphere from additional evaporation and subsequent condensation can, in principle, heat the tropical roposphere by several degrees but the actual temperature change observed is much smaller. What physical mechanisms act to limit the temperature change of the tropical troposphere?

The last three volcanoes large enough to significantly affect the atmospheric transmission of solar radiation have shown large differences in transmission and in mass of stratosphere aerosol yet the temperature changes produced in the stratosphere have been almost the same (5-6°C). What are the factors that act to limit the temperature change of the tropical lower stratosphere?

During an ice age, temperatures fluctuate over a fairly wide range as shown by oxygen isotope ratio in ice cores yet the lowest temperatures reached, corresponding to about -42% in $\delta^{18}O$ values in the Summit core in Greenland (GRIP, 1993), are usually the same. What are the mechanisms that act to limit these high latitude surface air temperatures?

These questions and associated ramifications are examined here.

Sea Surface Temperature Limits

The factors that control temperature change in the oceanic surface layer may be written:

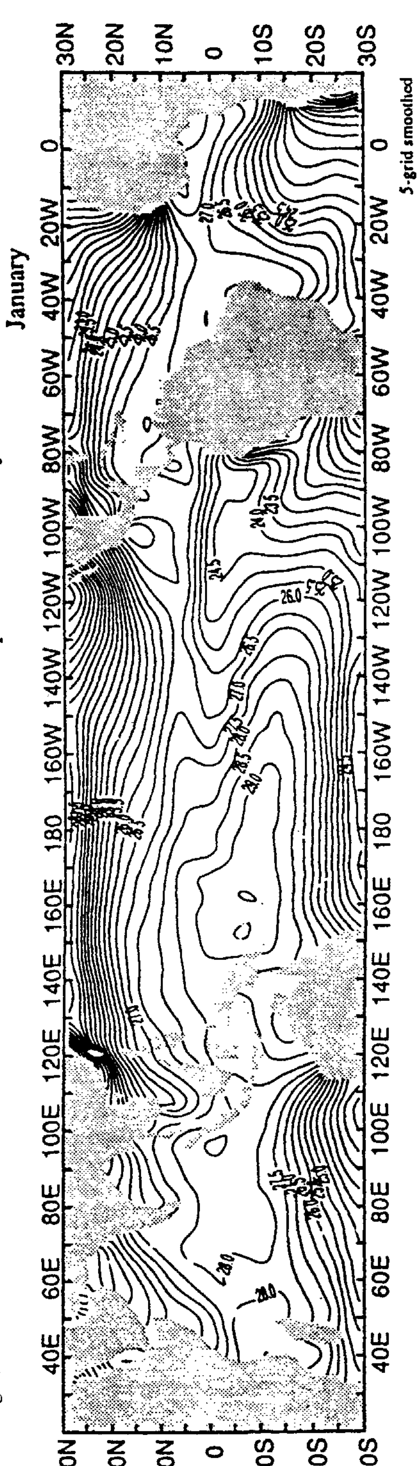


Fig 1(a)

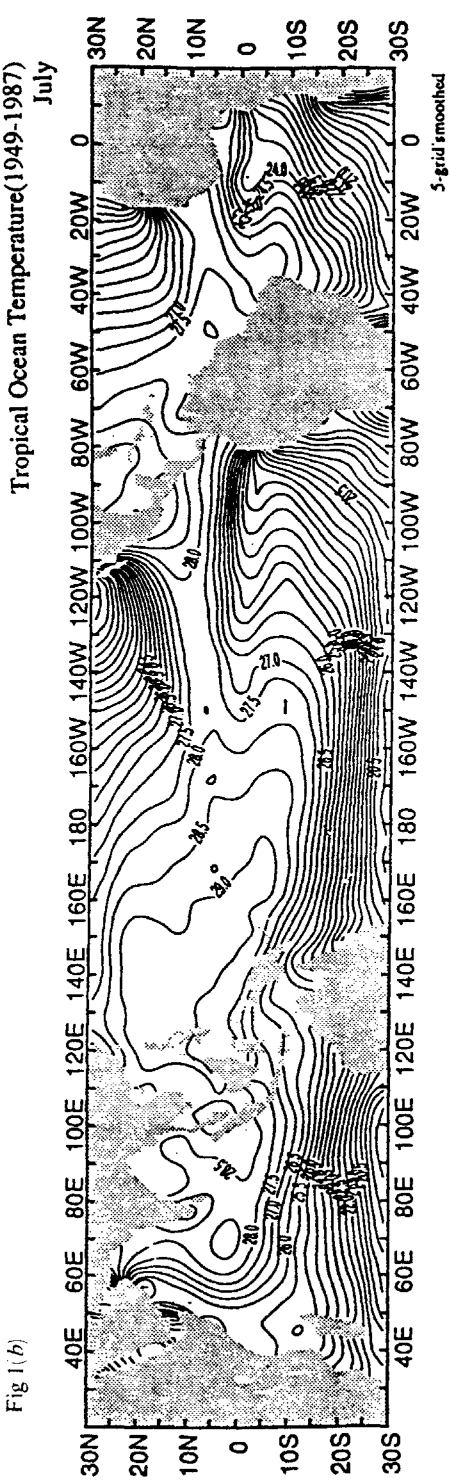


Fig 1(b)

Fig 1 Mean values of tropical sea surface temperatures (°C), 1949-1987: (*a*) January; (*b*) July

$$\frac{\partial T_s}{\partial t}=\frac{Q_{sol}-Q_{IR}-Q_{SH}-Q_{LH}}{\int_{-D}^{0}\rho C dz}-u\frac{\partial T_s}{\partial x}-v\frac{\partial T_s}{\partial y}-w\frac{\partial T_s}{\partial z}-\frac{\partial}{\partial z}K_z\frac{\partial T_s}{\partial z}, \quad \ldots(1)$$

where T_s is a surface layer temperature, u, v and w are oceanic flow components in the eastward, northward and upward directions (themselves denoted as x, y and z respectively) and K_z is a vertical eddy diffusion coefficient. The first four terms on the RHS denote the surface energy input by solar radiation Q_{sol}, the surface loss by infrared radiation Q_{IR}, the loss as sensible heat Q_{SH} and the loss as latent heat Q_{LH}. ρ represents the density of seawater and C the specific heat and D the thermal interaction depth over which the solar radiation is absorbed. T_s may be constant down to 200-300m in the west Pacific but only to 10m in the tropical East Pacific. One may in fact write the thermodynamic equation in terms of heat content changes and temperature of the isothermal layers but the same physics is involved. The fifth and sixth terms represent lateral advection while the seventh and eighth produce cooling by vertical advection (upwelling) or by mixing the surface layer with cooler water from regions below.

The surface energy components may be written as (Hsiung, 1986)[1]:

$$Q_{sol}=Q_0 Tr(1-A)(1-0.62c+0.0019\alpha), \quad \ldots(2a)$$

$$Q_{IR}=\varepsilon\sigma T_a^4(0.39-0.05\sqrt{e})(1-bc^2)+4\varepsilon\sigma T_a^3(T_s-T_a), \quad \ldots(2b)$$

$$Q_{SH}=\rho_a C_p C_H V(T_s-T_a), \quad \ldots(2c)$$

$$Q_{LH}=\rho_a L C_e V(q_s-q_a) \quad \ldots(2d)$$

and

$$Q_N=Q_{sol}-Q_{IR}-Q_{SH}-Q_L. \quad \ldots(2e)$$

Symbols are defined as follows:-

Tr, atmospheric transmissivity; A, albedo taken from Payne[2]; b, the cloud coefficient taken from Budyko's atlas[3]; c, cloud amount in tenths; α, solar noon altitude in degrees; ε, emissivity of water taken as 0.97; σ, Stefan-Boltzmann's constant; e, water vapor pressure; T_a, temperature of the air; T_s, temperature of the sea surface; q_a, specific humidity of the air; q_s, saturation specific humidity at T_s; ρ_a, air density; C_p, specific heat of air; L, latent heat of evaporation; C_h and C_e, transfer coefficients for sensible and latent heat taken from Isemer and Hasse[4] and Large and Pond[5]; and V, wind speed.

Typical values for eq. 2 are shown in Table I for the Eastern and western tropical Pacific. Eq. $2b$-$2d$ are all functions of surface temperature. While surface temperature does not appear explicitly in equation $2d$, q_s is a function of T_s being proportional to e_s/p, where p is surface pressure and e_s the saturation vapor pressure which is a function of temperature through the Clausius Clapeyron relationship. Typical conditions in the western Pacific warm pool are used to illustrate the dependence of the terms upon temperature in Fig. 2. The situation is simplified as the contributions by the second term of equation $2b$ are ignored. They are always very small (<10 Wm^{-2}) with the typical values of T_s-T_a that are observed (~ 1.0°C) (see Bottomley *et al.*, 1990)[6]. The mean

Table I

Components of the heat fluxes in different phases of El Niño events

	TWP (5S – 5N, 150-160E)		TEP (5S – O, 100 – 90W)	
	El Nino	La Nina	El Nino	La Nina
Latent Heat (W/M**2)	115	113	77	36
Sensible Heat	6	5	2	0
Long wave Radiation	44	47	47	48
Solar Radiation	195	206	200	204
Net Heat Fluxes (Ocean)	33	42	74	121
Mean SST (C)	29.1	29.3	25.1	22.1
Mean SST (C)	29.2		23.5	
Standard Dev. (C)	0.5		2.3	

Periods of each phase (year month)

El Nino:

1957.4-58.1 65.5-66.1 72.6-73.2 76.6-76.12 82.8-83.8 86.11-87.6

La Nina:

1964.4-64.7 67.9-68.1 70.6-71.2 73.11-74.3 75.10-76.1 78.3-78.10

1935.1-85.9

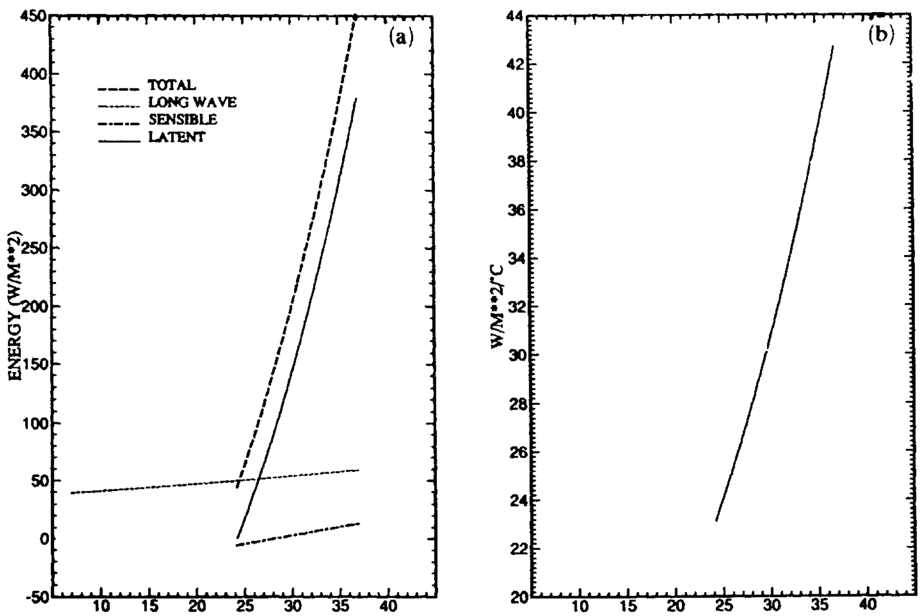


Fig 2 (*a*) Surface energy flux components for representative warm pool values ($v = 5$ m/sec; air temperature = 28°C; relative humidity = 80%); (*b*) sensitivity to temperature change wind speed is taken as 5 msec^{-1}, air temperature as 28°C and the atmospheric relative humidity is assumed to be 80%. For this region the temperature is often constant to 200-300m, as noted, so $\partial T/\partial z \rightarrow 0$ and, as is evident from the maps of Fig. 1, horizontal gradients are also very small. The first four terms of eq. (1) are then in approximate balance as observed. Fig. 2 shows that the equilibrium temperature is close to 30°C and that the sensitivity, obtained from the slope of the curve of energy versus temperature and shown in the right hand side of Fig. 2, is very low—typically it requires 30 Wm^{-2} to change the temperature by 1°C. This is thought to be due to the evaporational buffering introduced by Q_{LH} which is the largest term and therefore dominates this slope[7]. Fig. 2 is an updated version of similar graphs presented earlier[8,9]. It suggests that if additional radiative energy reaches the surface, for example by a decrease in cloudiness or an increase in atmospheric CO_2, it is offset in the warm pool region by additional evaporation. Note that the sensitivity of surface temperature to changes in incoming radiation would be much greater if infrared radiation provided the main balance during changes; the slope of the Q_{IR} curve is about 1 Wm^{-2}°C^{-1}.

Time series of sea surface temperature (SST) and the terms in equations 2 derived from the Comprehensive Ocean Atmosphere Data Set (COADS) are illustrated in Fig. 3 for the tropical east and west Pacific. The wind speeds used are included in Fig. 3*g*. It is clear that the sea temperature is quite stable in the tropical west Pacific showing little response to seasonal fluctuations in radiative input. The energy balance is between solar radiation and evaporational and infrared losses. There is reason to believe that the evaporational loss may be underestimated because when ships report calm winds there are often slight local flows accompanying convection that can encourage evaporation; approximately 25 Wm^{-2} may be added to Q_{LH} to take this process into account. As seen from Table 1 this reduces the net energy flux at the surface to close to zero (cf. Newell, 1986)[10]. A detailed consideration of this extra loss at low wind speeds has recently been given by Godfrey and Beljaars[11]. Conditions during El Niño are of considerable interest and it is clear from Fig. 3 that the waters of the tropical east Pacific receive less energy from the sun (because of more clouds—Fig. 3*b*) and lose more energy by evaporation during the 1982-83 El Niño than in the period immediately before and after (Fig. 3*d*). The water there warms towards the evaporational limit temperature (see Fig. 3*a*) of the tropical west Pacific[10]. In this region the upwelling terms are normally quite important and can account in general terms for the marked seasonal cycle[12]. But when upwelling and the other oceanic advection terms diminish as happens during El Niño the surface temperature in the east Pacific becomes more influenced by the surface flux terms[10] and rises towards the same limit that prevails in the west. Notice from Fig. 3*f* that the total surface flux tends towards zero in the 1982-1983 period, a condition normally more representative of the west Pacific.

The ocean surface temperature in the tropical west Pacific thus seems to be controlled by evaporational buffering with the same situation governing the maximum temperature achieved in the east Pacific during El Niño situations. By subtracting monthly mean values from 30°C, a maximum anomaly map for

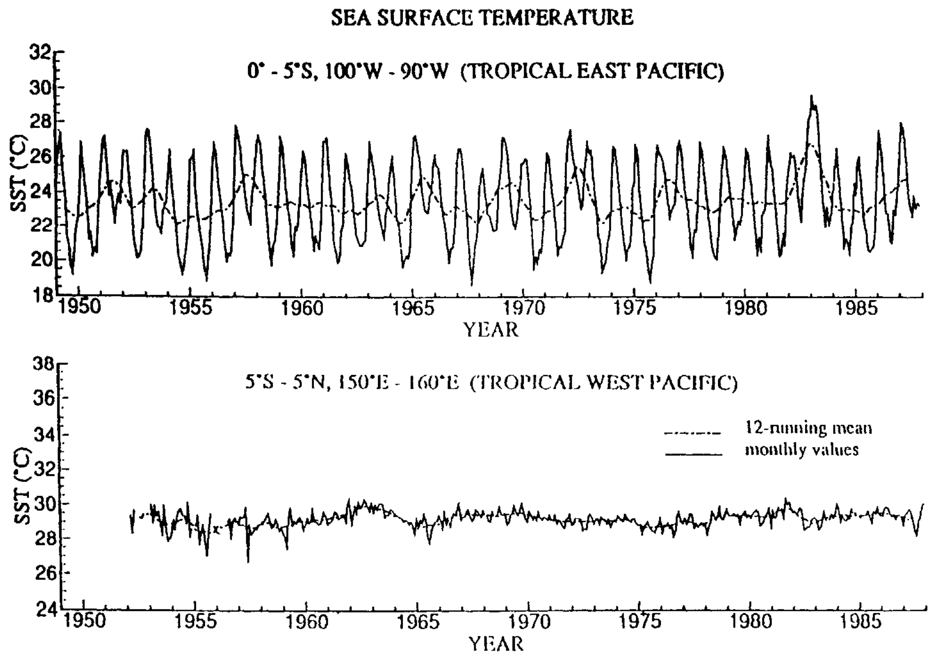


Fig 3(*a*)

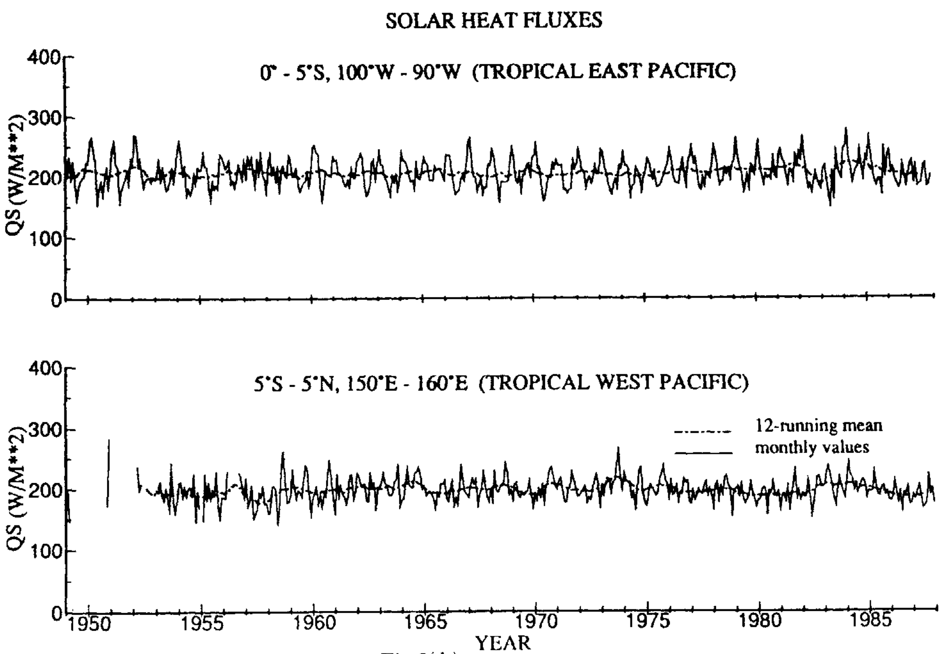


Fig 3(*b*)

Fig 3 Sea surface temperature, wind velocity and components of surface energy flux for the tropical east and west Pacific (*a*) sea surface temperature (*b*) incident solar radiation (*c*) sensible heat (*d*) latent heat (*e*) long wave radiation (*f*) net heat flux (*g*) wind speed. Long ticks are January [Figs 3(*c*), (*d*), (*e*), (*f*) and (*g*) are seen in the following pages]

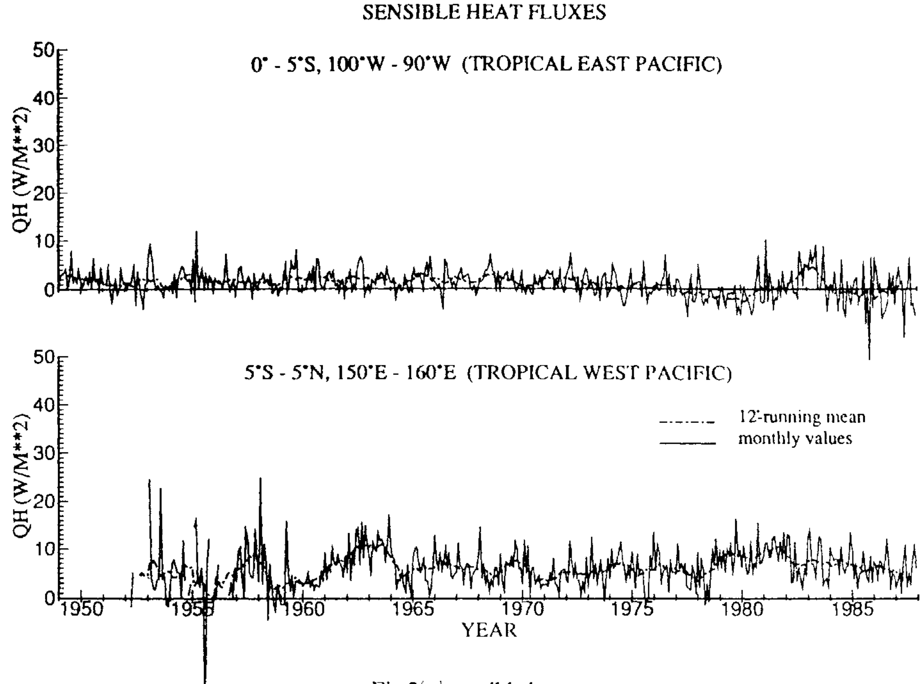


Fig 3(*c*) sensible heat

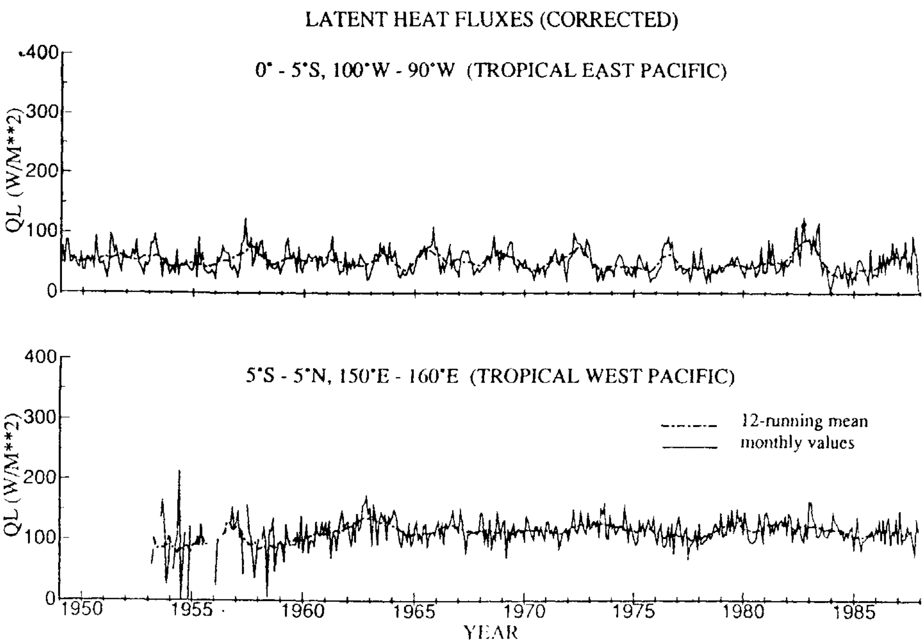


Fig 3(*d*) latent heat

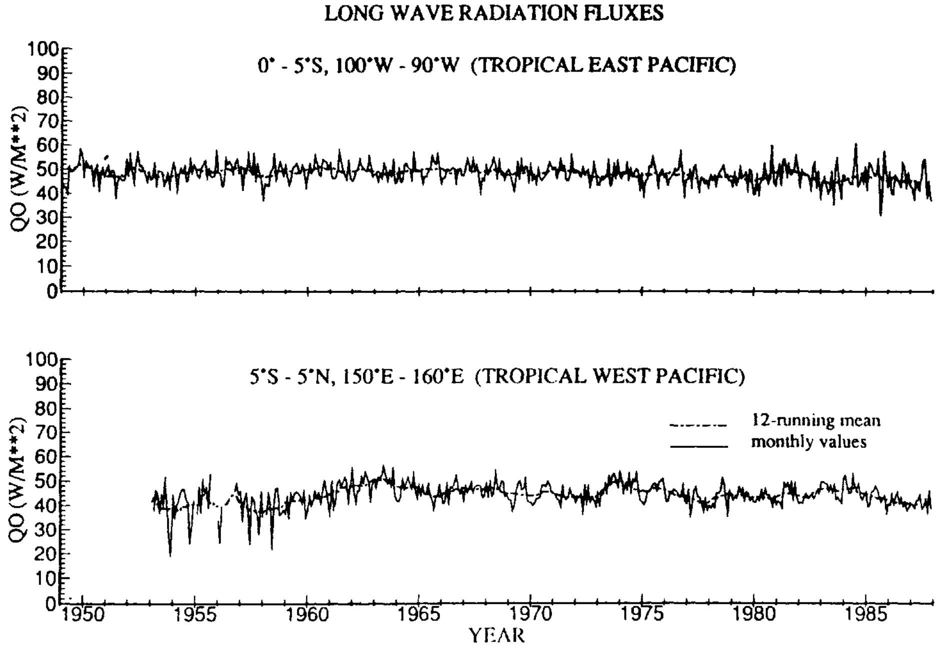


Fig 3(*e*) long wave radiation

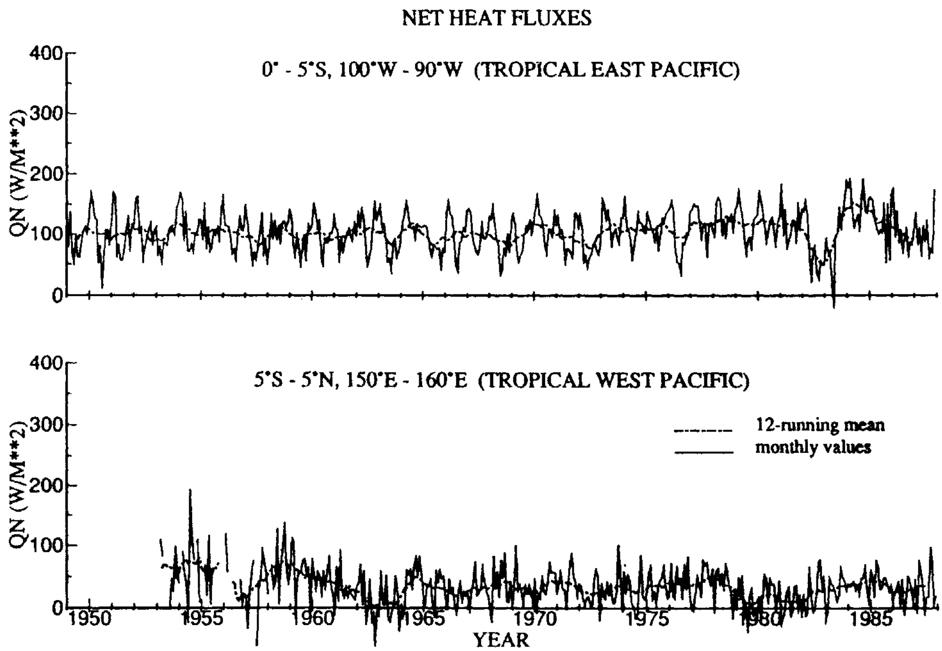


Fig 3(*f*) net heat flux

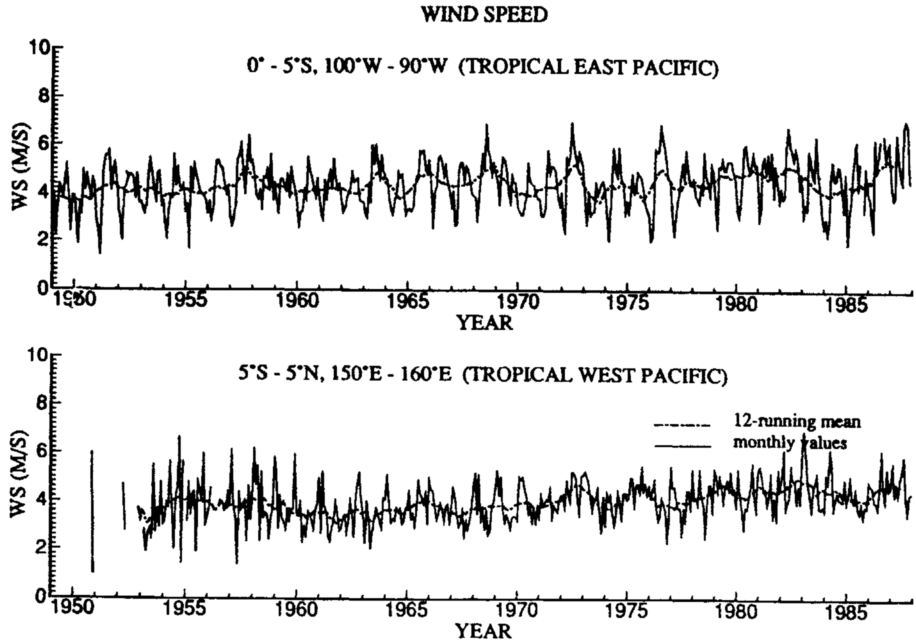


Fig 3(g) wind speed. Long ticks are January

each month can be reconstructed[13]; the actual anomalies that occurred in 1982-83 (up to ~6°C) are quite close to these maximum values. It is noteworthy that it is the Clausius Clapeyron relationship which controls the maximum temperature of the tropical ocean and therefore of the air in contact with the ocean, rather than the Stefan-Boltzmann law which is often invoked in simplified models. The evaporational buffering will also limit the minimum temperatures achieved in the warm pool region. Decreased solar radiation at the surface caused by greater than average cloudiness would be accompanied by decreased latent heat loss.

The limit on the minimum temperature in the tropical east Pacific is controlled by the balance between the net surface energy flux and the oceanic advective terms. Except during a strong El Niño the surface energy balance is about 100 Wm^{-2} (*see* Table I). The net contribution of the advective terms is quite complex; each is likely to be important. An excellent discussion of the energy budget of this eastern tropical Pacific region has been provided by Wyrtki[14]. Under the influence of a wind stress towards the west there are Ekman drifts away from the equator. These are accompanied by upwelling which brings up colder water from below which is then spread polewards by the Ekman drifts. In addition, the westerly stress builds up sea level towards the western Pacific and the consequent zonal sea level gradient has associated with it geostrophic drifts towards the equator thereby contributing to warm water advection. Zonal currents will also provide cold water advection in the east. We suggest here that as the wind increases there may be a minimum water temperature set by a balance between these advective terms. Stronger winds would increase upwelling and therefore cooling but at the same time increase the zonal sea level difference and thereby the equatorward geostrophic drift and the associated advective warming. It remains to work this hypothesis out with data; we have only examined the surface flux terms and the upwelling terms to show that their combination gives a reasonable explanation for the seasonal cycle[12].

A summary of the five terms in Fig. 3 divided into El Niño and La Niña phases is shown in Table I for the two regions. Incorporation of the factor to modify low wind speed situations as noted above adds about 25 Wm^{-2} to the latent heat loss and reduces the net flux in the west Pacific to close to zero, a condition invoked for Fig. 2. This condition would also be expected to apply to the east Pacific when SST there moves towards its equilibrium value during El Niño[10], an example of which appears in Fig. 3*f* in 1983.

Comparison of SST Evaporative Upper Limit with Other Ideas

Graham and Barnett[15] have also investigated warm pool SST limits and show that deep convection sets in above 27.5°C and that further SST increases have little effect on the intensity of the convection. Their work shows that large-scale divergence controls the deep convection at the higher temperatures. At the highest SST values convection is reduced according to their work, and this fits with our finding that in the Indian Ocean peak SST values (see histograms in Newell *et al.*, 1978)[7] occur in May when there is large scale atmospheric subsidence and divergence before the monsoon starts.

A different view was provided by Ramanathan and Collins[16] who suggest that highly reflective cirrus clouds shield the ocean from solar radiation and limit SST to less than 32°C. The problem with this explanation is that the warmest values appear where there is no cloud (see discussion below). If cloud is the limiting factor the SST should attain even higher values without cloud.

Wallace[17] argues that it is deep convection and not evaporation that provides the thermostat on sea temperature. He suggests that deep convection will rapidly carry away any surplus energy from "hot patches" and believes the skewness in the histograms of SST, with a sharp cut off at higher temperatures, is due to this effect. Ramanathan and Collins[18] have recently criticized Wallace's paper but seem to be mostly directing their remarks at evaporation as an explanation and not deep convection. They claim that evaporation decreases with increasing SST and that the evaporative heat flux decreases towards the warm pool. Both of these points are contrary to the computations of Table I or Fig. 3*d*. Waliser and Graham[19] have developed the work by Graham and Barnett[15] and they carry through the arguments in more detail, illustrating from data that the maximum temperatures occur under relatively cloud-free skies while the maximum of the SST population near 28°C is due to the reduction in the surface solar radiation by clouds associated with organized convection. They suggest that the thermostat is essentially the stability of the atmospheric column relative to large scale moist convection and solar flux. They imply that the cloud free limit is set by evaporation. Fu *et al.*[20] have recently joined the debate and from cirrus cloud observations suggest that cirrus cloud does not provide the main limit; they favour evaporation.

Tropospheric Free Air Temperature Limits

Studies of correlation patterns between global non-seasonal air temperature anomalies measured from space with microwave sounding units (MSU) and sea surface temperature anomalies show that over the last decade tropical air temperature changes throughout the entire tropical belt between about 25°N and 25°S are related to sea temperature changes in the tropical east Pacific; there is no relationship with the much smaller temperature changes in the tropical west Pacific[21]. The vertical extent of the atmosphere covered by the microwave sounder is roughly from the surface to 7km[22]. The lack of relationship between the two media in the west is thought to be due to the observation that the west Pacific is near to its evaporational limit; even though the actual evaporative loss is large, as is evident from Fig. 3*d*, the changes, unlike those in the east Pacific, are not systematically related to SST. The tropical air temperature changes by about 0.7°C per °C change of sea temperature in the tropical east Pacific according to these recent studies; this compares with previously reported values of about 0.5°C[13,23,24]. The tropical east Pacific is also thought to control global atmospheric concentration[7] of atmospheric CO_2.

The equation governing air temperature change may be written

$$\frac{\partial T_a}{\partial t}=\frac{Q_{RAD}+Q_{LH}+Q_{SH}}{\rho_a C_p}-\Gamma\omega-\frac{v\partial T}{\partial y}-\frac{u\partial T}{\partial x}, \qquad \ldots(3)$$

where eddy flux convergence terms, which are small in the tropics, are ignored. u, v, and ω now refer to atmospheric motion with $\omega = \mathrm{d}p/\mathrm{d}t$ and ρ_a is the density of the air. Γ is a stability factor $\frac{RT}{pC_p}-\frac{\partial T}{\partial p}$ where p is pressure and R is the gas constant.

The first question is—which of these terms is of most importance in producing the tropical belt of increased temperature that accompanies El Niño? From Fig. 3 we note that additional evaporation injects water vapor into the tropical belt. Using a model of the tropical atmosphere developed by Gill[25] the second author has modeled the tropical atmosphere for the case where additional latent heat of about 40 Wm^{-2} is injected into the eastern Pacific and finds that an increase of temperature of about the observed size (~ 1.0°C) occurs by adiabatic subsidence (the $\Gamma\omega$ term). Another possibility is that there is a change in the meridional heat transport from the tropics to middle latitudes; the appropriate formulation may be obtained from the last two terms of eq. (3) but the data necessary to study this possibility is not presently available here. The model produces a warm tropical strip but it seems that some other factor is involved to offset the adiabatic subsidence and limit the troposphere temperature increase.

The term Q_{RAD} incorporates at least five sub-components: heating by visible and near infrared (0.6μm-4μm) radiation absorbed by gases and aerosol and cooling by infrared emissions from water vapor, carbon dioxide and ozone. The tropospheric variable which changes most in the near infrared portion of the spectrum is water vapor. The additional water vapor given off by the tropical eastern Pacific during El Nino causes additional heating through enhanced near infrared absorption as well as condensation. But the heating is offset by additional infrared cooling which is also dominated by water vapor. Over most of the troposphere above the boundary layer increased water vapor is associated with increased infrared cooling rates[26]. The horizontal temperature gradients are relatively small in the tropics. Thus it seems likely that these radiative mechanisms, coupled with the latent heat liberation, may together produce another limit to atmospheric temperature rise. To study the role of changing water vapor in the east Pacific on the near infrared heating rates and the infrared cooling rates we have used a radiative heating rate program, designed by T Dopplick and described and revised by Hoffman[27], applied to wa-

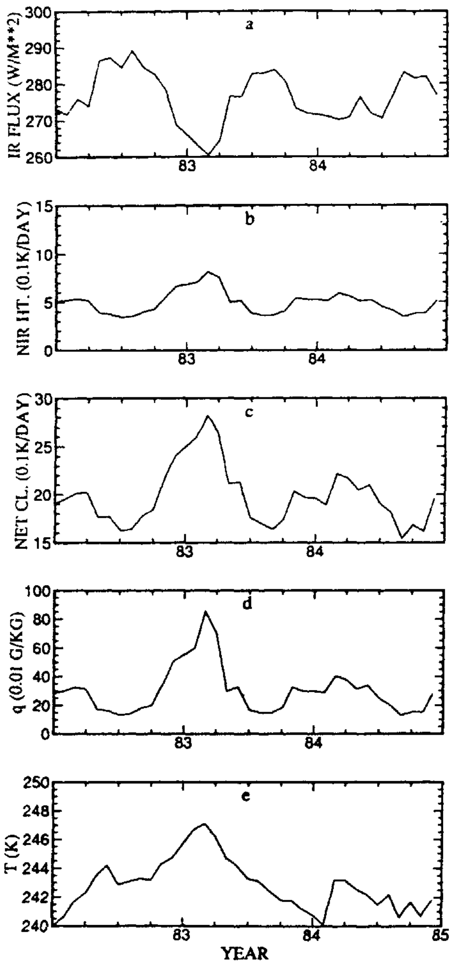


Fig 4 Heating rates for Atuona (9°40'S, 139°02 W) based on mean monthly radiosonde data: (*a*) infrared flux at 200 hPa; (*b*) heating by near infrared absorption at 300 hPa; (*c*) net cooling by infrared radiation at 300 hPa; (*d*) specific humidity at 300 hPa; (*e*) observed temperature variation at 300 hPa, 1982-1984. Long ticks are January

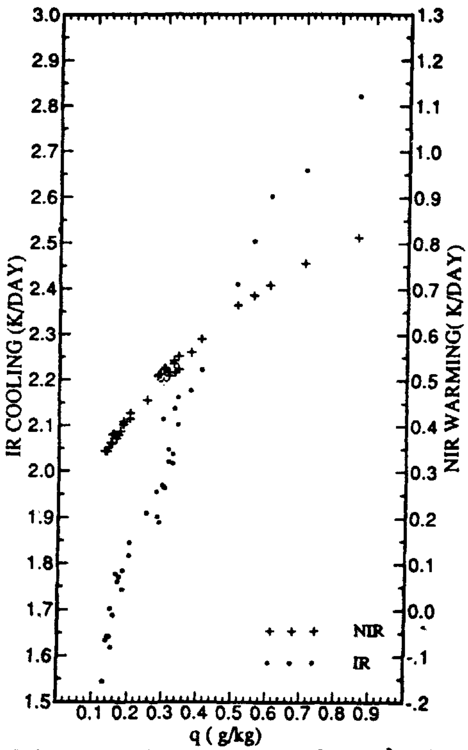


Fig 5 Infrared cooling and near infrared heating rates with water vapor mixing ratio at 300 hPa, based on 1982-1986 period

ter vapor and temperature values taken from monthly mean soundings for Atuona (140°W, 10°S) for the 1982-84 period. These calculations, performed for another purpose, assumed cloud-free skies (M. Lozano, *private commun.*). Values of temperature and specific humidity at 300 hPa are shown in Fig. 4 together with the infrared cooling, the near inf. .red heating and the infrared upward flux at 200 hPa. As already noted when the eastern tropical Pacific warmed in the 1982-83 El Niño the specific humidity increased and with it the near infrared heating and the infrared cooling. As expected the upward infrared flux also decreased. Obviously there would be concomitant changes in the horizontal energy fluxes to be taken into account if we were trying to balance the energy budget for the tropical belt; here we concentrate on the local cooling rate.

The near infrared heating rates and total infrared cooling rates for 300 hPa are shown in Fig. 5 as a function of specific humidity. At a value of about 0.5 g kg^{-1} the sensitivities of these rates to changing water vapor are 0.6°C day^{-1}/g kg^{-1} and -1.8°C day^{-1}/g kg^{-1} respectively. The infrared cooling dominates and provides a buffering process for the temperature changes. The additional evaporation and rainfall is not known very well and needs careful study before this temperature limit can be established. It is important to recognize that during El Niño, when additional evaporation from the eastern

Tropical Pacific occurs, the condensate can be carried by the zonal winds throughout the tropical belt and also to higher latitudes even though the largest water vapor anomalies occur in the eastern Pacific. The heating rate from the additional condensation can be quite large (~ 0.5°C day^{-1}) but the full value is not realized because of inrared cooling. In general, the heating by condensation occurs over a region with a smaller latitude span that the region in which heating by subsidence occurs. In order for a net temperature increase to occur the over subsidence heating must outweigh the additional radiative cooling or the meridional energy flux to higher latitudes must decrease.

In the seasonal cycle water vapor has a semi-annual cycle in the equatorial zone[28] and the tropical microwave temperatures also show a semi-annual temperature maximum in the same phase (W. Hu, *private commun.*) which may be produced by near infrared absorption. The additional heating during El Nino will thus be composed of condensation heating from rainfall, near infrared solar absorption, and warming by subsidence. Numerical values of a possible limit remain to be worked out from a more comprehensive set of grid point data which includes clouds.

Lower Stratospheric Temperature Limits

Another region where a natural temperature limit may exist is in the tropical lower stratosphere. This region is sometimes invaded by volcanic aerosol and

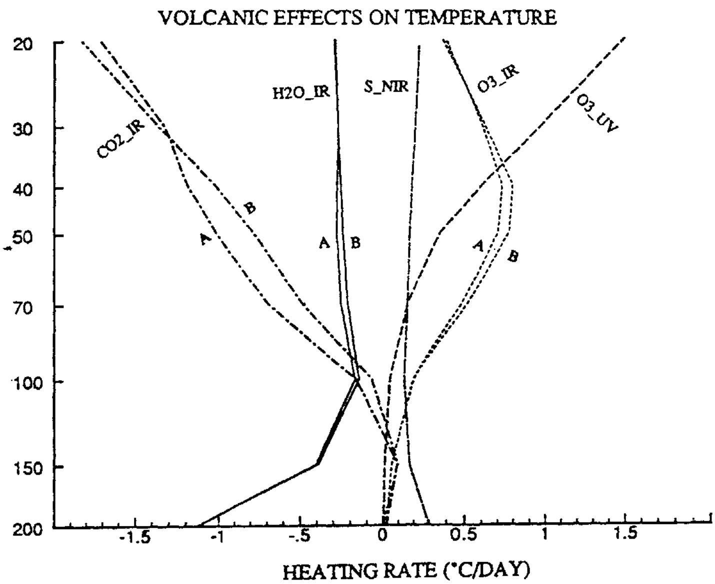


Fig 6 Radiative heating and cooling rates for tropical radiosonde sounding for Guam for conditions before (*B*) and after (*A*) the eruption of Mt. Pinatubo in June, 1991. S NIR is near infrared heating by water vapor, molecular oxygen and carbon dioxide; H_2O IR and CO_2 IR represent infrared cooling by these gases, O_3 UV is heating by ozone absorption of solar radiation, and O_3 IR represents infrared heating. At 50hPa, A temperatures are higher than B by 6°C

absorption by this aerosol of sunlight and perhaps, in some cases, of upwelling infrared radiation causes temperature increases. According to radiosonde data, these measured about 6°C after the Mt. Agung eruption[29] in March 1963 and about 5°C after El Chichón[30,31] in April 1982. Recent estimates based on MSU data (Christy, *private commun.*) show that the Pinatubo and El Chichón eruptions gave similar tropical stratosphere increases. The amounts of dust in the stratosphere from those eruptions has differed widely with Pinatubo thought to produce the largest amounts. It is therefore suggested that there may be another temperature limit in this region. Using the same radiation formulation as that discussed above, it can be seen from Fig. 6 that atmospheric CO_2 in the tropical lower stratosphere produces cooling rates at the higher post-volcanic temperatures (the difference was assumed to be 6°C at 50 hPa) which tend to offset heating by aerosol absorption[32]. The infrared cooling from CO_2 increases with temperature and in this case it is essentially the Stefan-Boltzmann law that is controlling the limit. It is the component of the first term of equation (3), Q_{RAD}, that provides the balance as seen in Fig. 6, although changes in vertical motion may also be involved.

Possible Limits on Ice Age Temperatures

How do these limits fit with what is known about conditions during the past ice age? Temperatures in the tropical eastern Pacific were lower than the present by 4-6°C at 18000 B.P. in the original estimates[33] but only about 2°C in more refined estimates[34]. These values need more work before they can be compared with present values of seasonal SST changes such as those in Fig. 3*a*. Pollen data from Africa[35] and South America[36] show descent of the tree line of 1000-1100m and 1200-1500m respectively in these two regions, corresponding to temperature decreases of more than 6°C. Because temperatures were generally lower the atmosphere was able to hold less water vapor. Specific humidity values for the past ice age were scaled from present values with the assumption that the early CLIMAP surface temperature results were valid and that relative humidity remained the same; temperature lapse rates were also taken to be the same as present values. Radiative cooling rates were computed for these synthetic ice age conditions and the assumption made that the first three terms on the right hand side of equation 3 summed to zero[37]. It turned out that the cooling rates decreased by about 10% relative to present values and the conclusion was made that the latent heat term and therefore rainfall must also have decreased by this amount. This point was in reasonable agreement with ice age data (e.g. Flint, 1971)[38] and with findings from general circulation models (e.g., Gates 1976 found a decrease of 14%)[39].

We have argued already that the tropical west Pacific warm pool temperature is essentially stable and that the tropical east Pacific surface temperature is limited on the lower side. Similar arguments may be made for the warm regions in the Indian Ocean and the upwelling in the Atlantic. The implication is that events in middle or high latitudes must be responsible for the lower limit on ice age temperatures. Latent heat liberation is the main factor in eq. (3) providing energy to the atmosphere. As noted above, when the atmosphere cools this term diminishes. But as high latitude cooling increases meridional temperature gradients steepen and overall wind velocities increase[40]. At present largest evaporation occurs off the east coast of the continents in winter as cold dry air streams out over the ocean. With stronger winds and ice present on the continents throughout the year in the ice ages, there will most likely be a minimum value for this evaporation term and therefore a lower limit on Q_{LH} and temperature in eq. (3).

This possible lower limit may be treated from another perspective by considering the meridional fluxes of energy by the atmosphere and ocean. When one or both of these fluxes diminishes below its long term value there is a resulting energy deficit at high latitudes and ice may build up; conversely the ice would diminish when the fluxes are above present values. The change in the amount of ice in the polar cap regions can be used to assess the associated change in he fluxes[41]. The time assumed for ice age build-up in our previous work was about 8000 years. A more modern view may yield a higher value by about a factor of two and this gives a value for the required flux changes of about 2.5×10^{14}W. Present day fluxes are shown in Fig. 7. At 30°N atmospheric fluxes (taken here from Oort and Vonder Haar, 1976)[42] are a maximum in the northern hemisphere winter while oceanic fluxes (taken here from Hsiung *et al.*, 1989)[43] are maximum in northern hemisphere summer. The radiation budget which these fluxes balance and the relative roles of heat stored in the ocean and meridional energy fluxes in maintaining winter temperatures is discussed elsewhere[8]. Our previous suggestion was that a decrease in the oceanic energy flux initiates a change towards an ice age[41]. A little more is now known

(*a*)

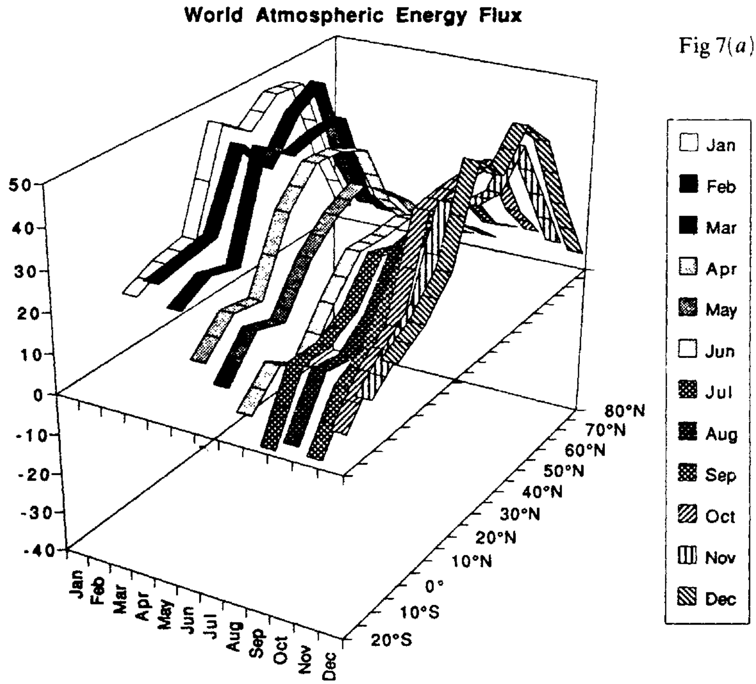
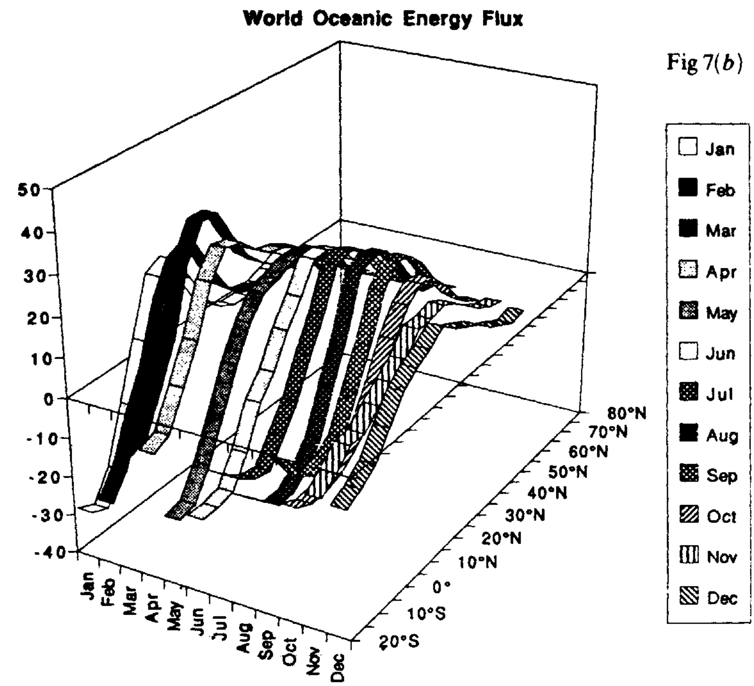
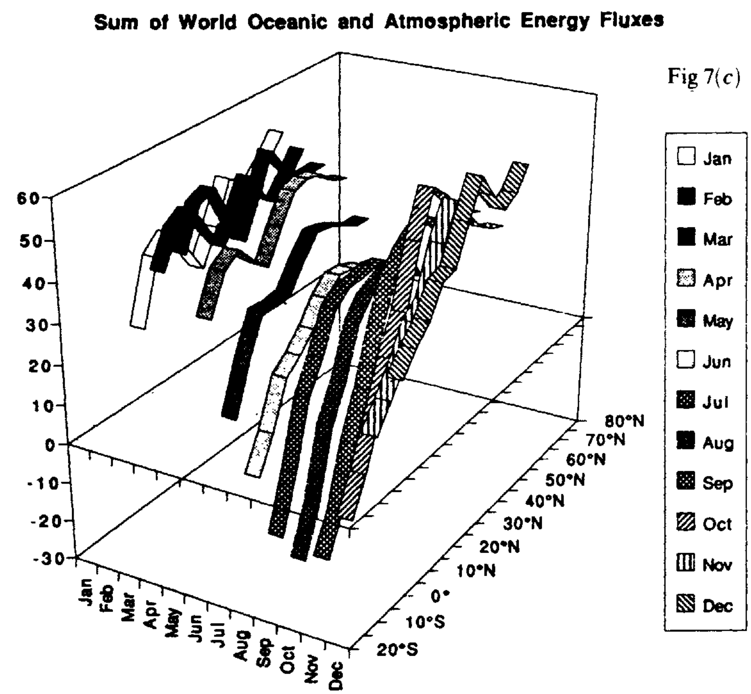


Fig 7 Meridional world energy flux (units 10^{14} W): (*a*) atmosphere;·



(b)



(c)

Fig 7 Meridional world energy flux (units 10^{14} W):
(b) ocean; (c) sum of atmosphere and ocean

about the annual cycle of the oceanic fluxes for each ocean separately (Hsiung *et al., loc. cit.*)[43] and about the possible linkage of the Indian and South Atlantic Ocean (Gordon, 1986)[44]. It is thought that the Indian Ocean supplies energy to the South Atlantic which then crosses the equator into the North Atlantic. Thus the place to look for changes in solar radiation that may influence ice formation is not necessarily high latitudes in summer but may be tropical latitudes whenever there is an energy transport from the Indian Ocean into the Atlantic. The breakdown of energy flux components appears in Figs 8. Clearly the Indian Ocean carries energy southwards and the Atlantic northwards in April, July and October but the reverse is true in January. This linkage may be the best place to look for the influence of Milankovitch solar radiation variations on climate. If the oceanic energy supply to the North Atlantic is the ultimate mechanism that initiates ice ages, the fact that the atmospheric energy flux increases with the meridional temperature gradient (*see* Newell, 1974 and Fig. 7*a*)[41] suggests that at some point the atmospheric flux compensates for the

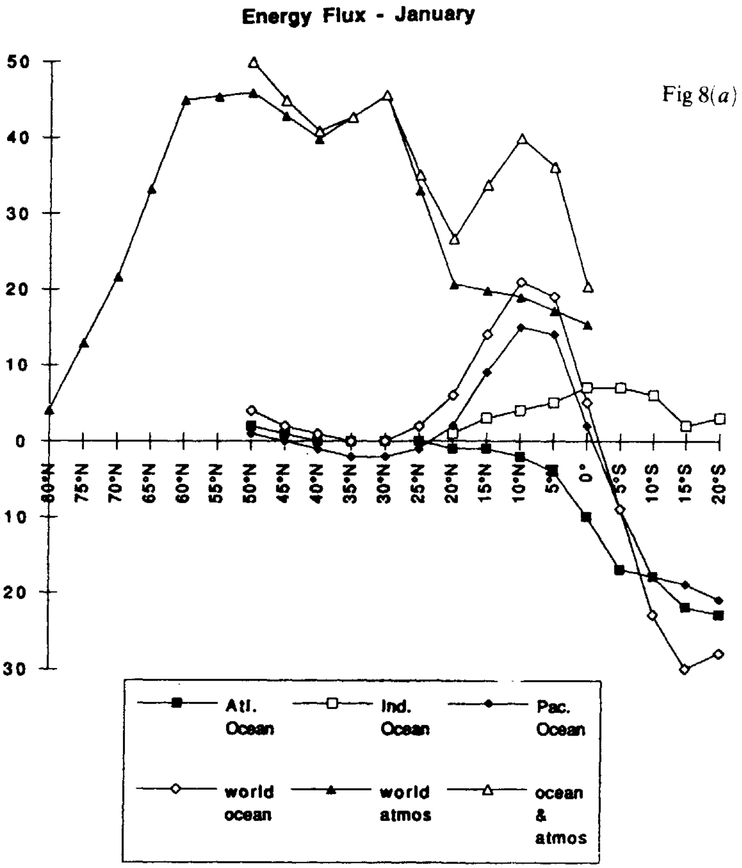


Fig 8 Individual components as indicated of meridional energy flux (units 10^{14} W): (*a*) January;

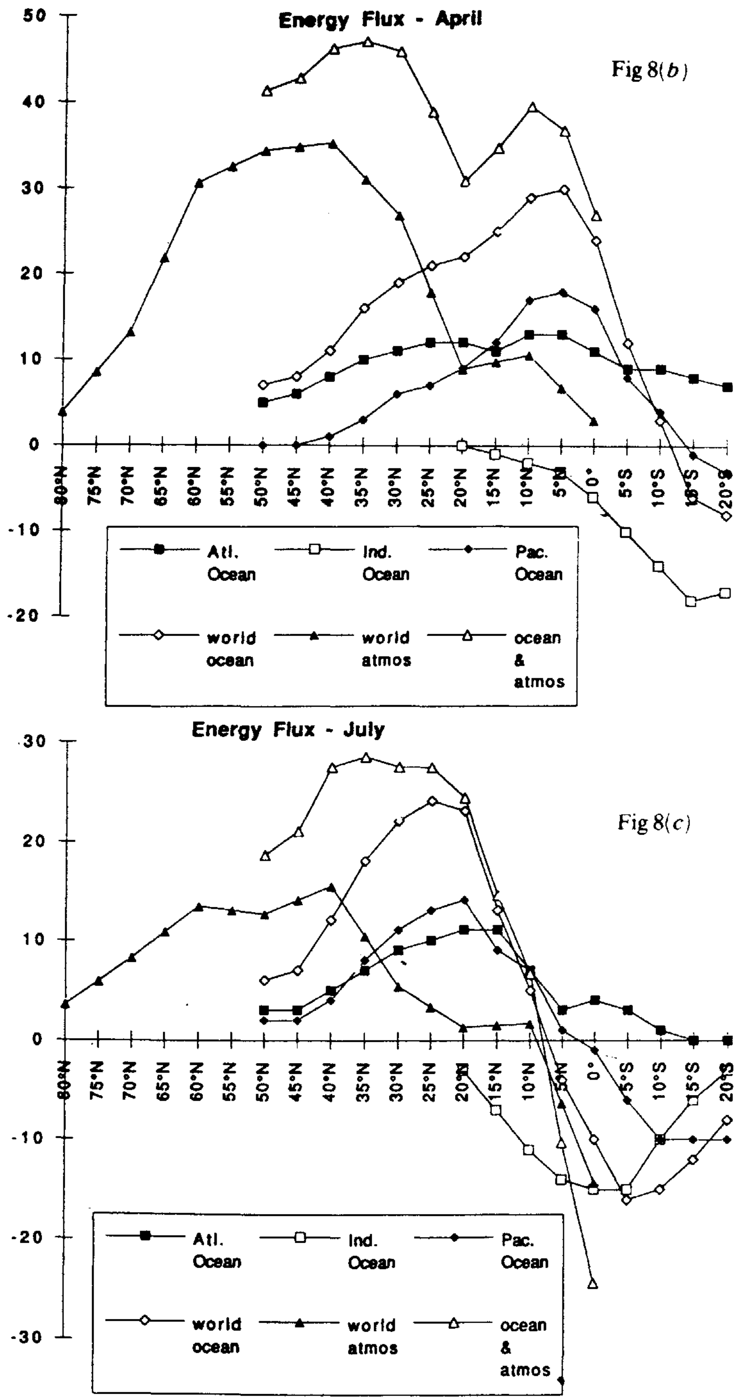


Fig 8 Individual components as indicated of meridional energy flux (units 10^{14} W): (*b*) April; (*c*) July

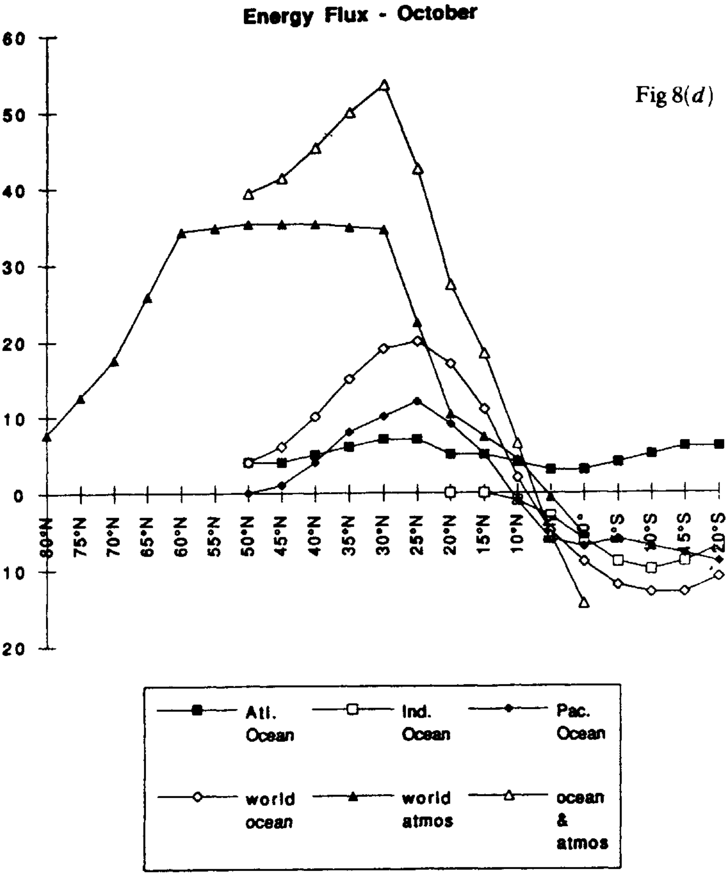


Fig 8 Individual components as indicated of meridional energy flux (units 10^{14} W): (*d*) October

reduction in oceanic flux and the overall energy budget returns to present values. This would have happened at about 18,000-14,000 years BP at the end of the last ice age and is the time at which the oceanic conveyor belt is thought to have "switched on"[45]. It is easy to see that there could have been a net meridional energy flux surplus at this time, quite apart from any Milankovitch variations, so that ice melting at high latitudes would occur. The presence of the ice maintains this steep meridional temperature gradient for some time. The value we deduced from the ice volume change observed between 18000 years BP and the present was about 2.5×10^{14}W—a relatively small number when compared with present day values given in Fig 7*c*.

Conclusions

The issue of temperature limits in various parts of the climate system has been discussed.

The tropical west Pacific warm pool is kept near to 29-30°C by what may be termed evaporational buffering. It is basically the Clausius Clapeyron relationship that governs this limit; the appropriate sensitivity is close to 30 $Wm^{-2}°C^{-1}$.

The tropical east Pacific has the same upper limit. The minimum temperatures there, it is suggested, are controlled as a balance between upwelling and associated Ekman divergence, surface heat flux and geostrophic motion towards the equator giving horizontal advection and being controlled by the westward wind stress.

Tropical free troposphere temperatures are thought to be limited by extra water vapor infrared cooling balancing extra near infrared heating, extra condensation and associated adiabatic subsidence, all accompanying El Niño.

Tropical lower stratospheric temperatures rise during volcanic activity. Heating by aerosol absorption is offset by infrared emission from CO_2 which increases at the higher temperatures so that maximum rises appear to be about 6°C.

Ice age minimum temperatures are limited by meridional energy transport by the coupled atmosphere and ocean.

These issues represent a selection of ideas suggesting that temperature limits occur at various points in the climate system and are consistent with physical principles that must be quantitatively understood before we can claim a first order understanding of climate.

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