The weather and climate of the tropics

Part 1 – Setting the scene

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In this series, I will describe various aspects of tropical weather and climate as it is understood today. The series will cover these topics: the sub-tropical jet streams; synoptic-scale weather systems; forecasting significant cloud and associated weather; dry environments; the development and maintenance of monsoon systems; tropical revolving storms; and mesoscale convective complexes. Each part will include a description of the effects of tropical weather.

This series is not intended to be a detailed mathematical description of tropical weather. A number of very good texts are available for those who wish to go into greater depth; in particular, I recommend Hastenrath (1991) and Riehl (1979).

Much of what we know about the weather has been focused on mid-latitude weather systems: first, because most early researchers came from Western Europe and eastern North America; and second, because of the risks and consequences of weather systems prevalent in these zones. However, although there are simple non-scientific descriptions of weather events from the tropics going back hundreds of years, it is only since the late 1960s that much scientific research has been carried out within the tropical zone. What we know of the weather (and, to some extent, the climate) of the tropics remains limited and has typically focused on severe weather events, such as tropical revolving storms (e.g. Emanuel, 2005), or data from a limited range of observing stations. However, many factors of the day-to-day weather are important in the tropics, not least for aviation and public safety.

For instance, the primary purposes of forecasting for aircraft operations in the tropics are safety and maximization of efficiency for the benefit of passengers and aircraft operators. The most accurate and appropriate forecasts will achieve this goal, using a mixture of numerical weather prediction products, observed data and good forecasting knowledge. It is the effects of the weather, in other words its outcomes, which must be considered.

The research carried out as part of the World Climate Research Programme since the early 1970s is very important in allowing us to understand many of the processes and associated weather of the tropical zone (Gates and Newson, 2006). Knowledge continues to grow through more recent research programmes, such as TOGA, which investigates the important links between the tropical ocean and the global atmosphere (Fleming, 1986). It is clear that the tropics have an important effect on weather systems throughout the globe, providing much of their energy.

How can we define the tropics?

The most commonly used definition of the tropics is the zone within which the Sun is directly overhead at some time during the year, i.e. the zone between the tropics of Cancer and Capricorn (23.45°N and 23.45°S, respectively). However, a more meteorological definition, rather than the elevation of the Sun at midday, is probably more useful to both the weather forecaster and the climatologist.

A simple method is to divide the globe into tropical and extra-tropical zones. This is a method often employed for the verification of numerical forecasts (WMO, 1982; Fuller, 2004). One such method divides the globe into two equal halves – tropics and extra-tropics – with somewhat arbitrary dividing lines at 30°N and 30°S. Conveniently, this latitude range includes almost all of the climatic zones that can be regarded as tropical: humid equatorial, savannah, semi-desert scrub and hot desert (Figure 1).

However, the zones of predominantly westerly winds make incursions equatorward of these lines of latitude, particularly in winter. In order to keep within a zone of

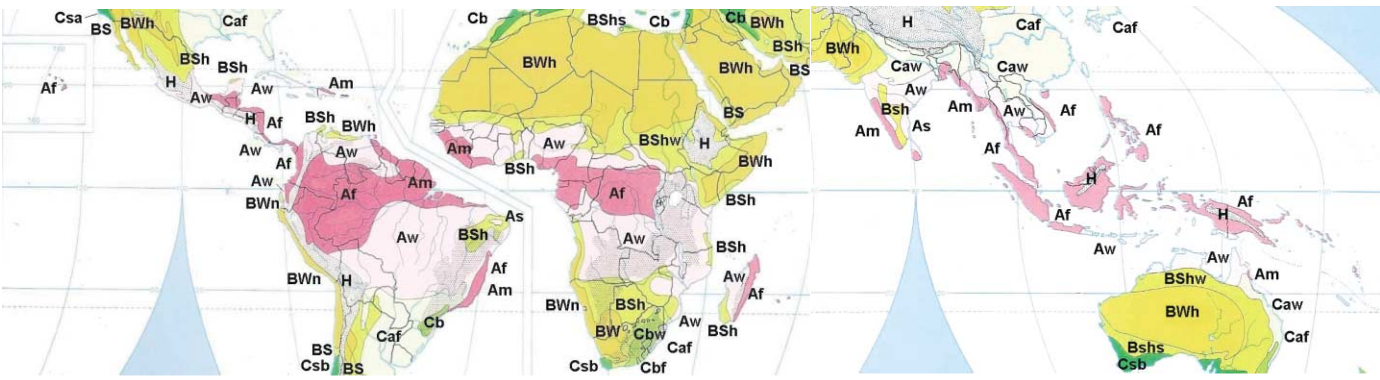

Figure 1. Climatic zones of the tropics: Af, Am – tropical rainforest; Aw – savannah; BSh – tropical steppe; BWh – tropical and sub-tropical desert; Cs – dry-summer sub-tropical; H – highlands; Cf – warm temperate with no dry season; Cs – warm temperate with dry summer; Cw – warm temperate with dry winter; stippled– modification due to altitude. The effect of high ground has a profound influence on climate in the tropics, particularly above about 2 km. (Using the system devised by Wladimir Koppen.)

predominantly easterly winds at most levels, a narrower zone must be used. Where an automated system, such as the verification of numerical forecasts, is not required, there are better climatological or synoptic-meteorological ways to define the tropics.

A more useful definition for the climatologist is based on the small annual variation of climate typical in the tropics. Riehl (1979, Ch. 2) proposed the definition as the area within which the diurnal temperature range exceeds the range of annual mean temperature. This has great value, since data can easily be sorted using this definition. However, the weather forecaster usually needs something more closely related to the daily weather, without reference to seasonal variations.

Using current weather, it is possible to define the tropical zone on a daily basis. The temperature difference between the tropics and middle latitudes causes a jet stream to develop at the poleward limit of the tropics. This sub-tropical jet stream (STJ) has a core close to 30°N and 30°S and has little high-amplitude wave development along it (Figure 2). The area between these jet streams has a tropospheric depth characteristic of the tropical zone. This depth allows us to define the periphery of the tropics, even when the STJ weakens or is absent, as often occurs in summer, as is usually the case in the Northern Hemisphere in summer. Furthermore, the STJs coincide with the transition from westerly lower tropospheric winds on their poleward side to the easterly Trade Winds on their equatorward side.

Although use of the STJ as the northern and southern limits of the tropics means that the tropical zone extends north of 40°N over Asia during the northern summer, it is appropriate, since the air to the south of it retains tropical characteristics. To provide consistency throughout this series of papers, some areas poleward of the mean latitude of the STJ will be discussed, since these areas spend part of the year within the meteorological tropics. This also allows inclusion of areas frequently affected by upper-tropospheric troughs in the STJ.

The tropical troposphere

The height of the tropopause varies little in the tropics, but is related to the mean temperature of the troposphere. Thus the tropical zone effectively contains a single air mass. However, variations in mean temperature and dynamics cause some variation and the highest tropopause heights are generally found close to the Equator. The height of the troposphere rarely extends above 17 km* and is most often between 15 km and 16 km. Height gradients are usually small, but increase somewhat near the sub-tropical jet streams. Indeed, the association of the STJ with a tropospheric depth of about 15 km provides a definition of tropical air.

* Altitudes given in this series of articles approximate the true heights. In the tropics, these are about 7% greater than the ICAO heights corresponding to pressure levels.

Despite its relatively uniform depth for much of the year, some variation occurs with the changing of the seasons and these variations are notably marked in the Northern Hemisphere summer. Between late May and late September, the tropopause is higher over north Africa and south Asia than it is close to the Equator. The intense warming of these land masses causes the troposphere to expand and the tropopause occasionally reaches a height of 18 km over Tibet. Smaller expansions occur over Australia, Africa and South America during the southern summer. The expansion is a key element in the development of the summer monsoon circulations (to be described in Part 6).

Associated with the great depth of the tropical troposphere is a high total (1000–500 hPa) thickness. This generally has a minimum around 580 decametres along the northern and southern boundaries of the tropical zone and may reach 590 decametres or more in places.

Winds in the tropical zone

Within the tropics, winds are often relatively light, in particular at upper levels. Over the Atlantic and much of the Pacific, these are westerlies throughout the year. Over the western Pacific, Indian Ocean and Africa, there are high-level easterlies close to the Equator (Figures 2 and 3). At low levels, Trade-Wind flows predominate, originating in the sub-tropical high-pressure systems (the areas of the 'doldrums') centred near

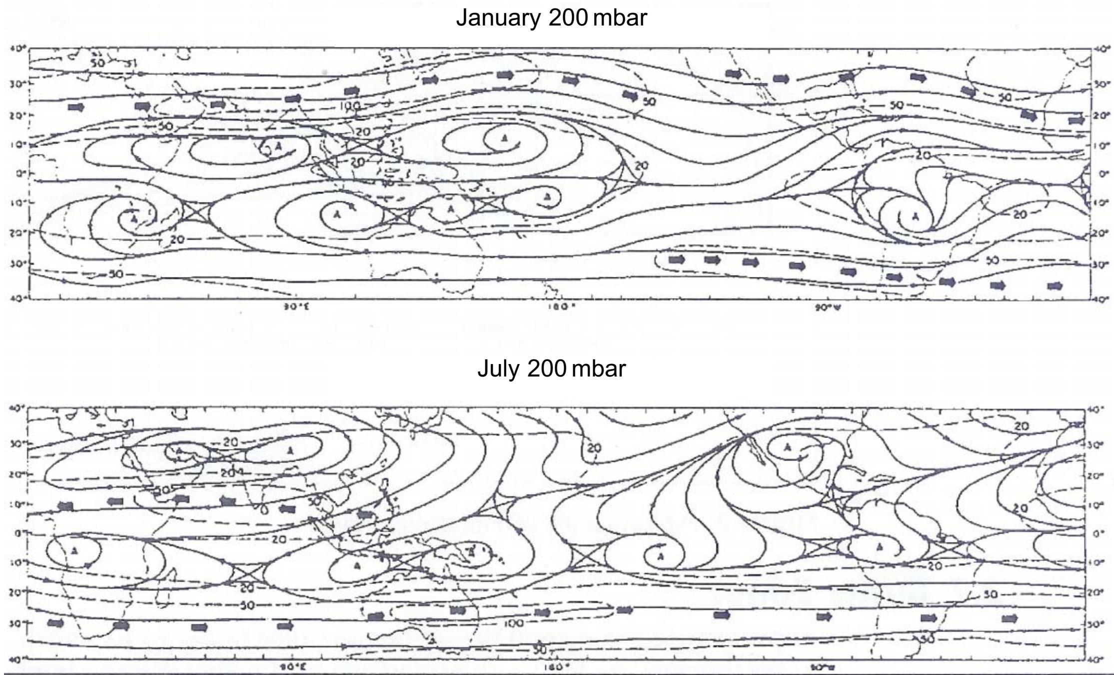

Figure 2. Mean annual wind flow (streamlines) at 200 hPa (c. 12 km) in the tropics with vector-mean isotachs (ms^{-1}) shown by dashes. The mean core of the sub-tropical jet streams is shown by arrows. The effect of using gridded vector-mean speed, as in this case, is to reduce the apparent speed of the winds, which are instantaneously stronger at all times and in all seasons. (The use of the 200-hPa also omits the strongest winds observed near the Equator, which develop between the 150-hPa and 100-hPa levels; these are indicated by arrows.) The easterly and westerly branches of the Walker circulation can be seen close to the Equator. (Adapted from Godbole and Shukla, 1981.)

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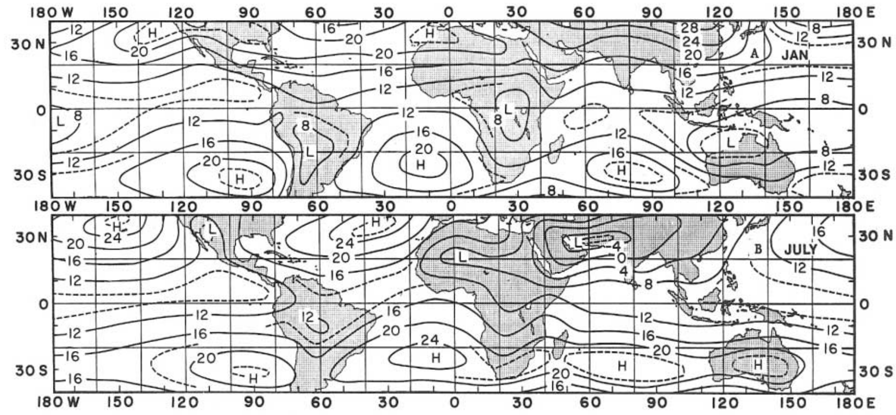

Figure 3. Mean sea-level pressure in the tropics (a) January, (b) July. Over high ground, in particular the Himalayas, mean sea-level pressure has no meaning, since the surface wind flow is influenced by pressure differences at higher levels. (Adapted from Godbole and Shukla, 1981.)

30°N and 25°S. North-easterlies in the Northern Hemisphere converge near the Equator with south-easterlies in the Southern Hemisphere providing the additional forcing necessary for deep convection within the tropics, forming a belt of convective cloud: the inter-tropical convergence zone (ITCZ). However, significant reversals in the low-level wind flow occur over and around the tropical continents during the change from winter to summer.

Summer monsoon circulations have westerlies at low levels, but middle- and upper-tropospheric easterlies strengthen as the troposphere warms and deepens to form an upper high over the continents, on the equatorward rim of which there is a steep temperature and height gradient. The greatest strengthening occurs as the equatorial high migrates away from the Equator and deepens. Close to the Equator, these winds may reach jet stream strength in a shallow layer near 15 km altitude. This jet stream is found only about 1000 m below the tropical tropopause. During the Northern-Hemisphere summer, these winds stretch all the way from South-East Asia to Africa's Gold Coast, although the main activity and highest speeds are generally across the southern tip of India, where speeds occasionally reach 60 ms^{-1}* and there is a local minimum over eastern Africa (Hastenrath, 1991). A similar, but weaker, jet stream forms over New Guinea and Indonesia in response to the warming of Australia during the southern summer.

The easterly winds at high levels diverge north and south away from the upper high that forms close to the Equator. North and south of 15° latitude, they become westerlies (Figure 2). The upper westerlies may reach jet stream strength to form the subtropical jet stream (STJ) along the poleward edge of the tropical air mass, close to 30°N and 30°S. The STJ is present throughout the year in the Southern Hemisphere, but has large speed changes between winter and summer in the Northern Hemisphere. In the northern winter, its speed may reach 110 ms^{-1} or more over east Asia and the western Pacific. However, its speed is rarely more than 50 ms^{-1} in high summer with a mean closer to 25 ms^{-1}.

At medium levels, winds of moderate strength often form wave trains, known as easterly waves, which form in response to temperature differences and convective development. These have a strong association with severe weather, notably summer mesoscale weather systems over India, and the squall lines of Africa and parts of the Pacific Ocean (Atkinson, 1971; Leroux, 2001). Over the Atlantic, easterly waves are intimately associated with the development of tropical revolving storms (Emanuel 2005). The low value of the Coriolis force close to the Equator presents a difficulty in the assessment of wind speed and direction, so streamline analysis, rather than conventional pressure analysis, is generally used in the tropics, as described in Box 1.

The weather patterns and climates of the tropics

The tropical region experiences only gradual changes in weather patterns and variations are generally small, even between seasons. The main changes are between dry and wet seasons, marked (i) by the northward and southward movement of the inter-tropical convergence zone (ITCZ) in the central (equatorial) portion and (ii) by the winter incursion of cooler air at altitude near the poleward extremes. Even with this movement, there is almost no seasonal weather fluctuation within about 5–10° latitude of the mean position of the ITCZ. However, since most of the world's hot deserts have their equatorial flank within 20° of the Equator, marked by transition to savannah vegetation (e.g. the Sahel of West Africa), seasonal variations can be large close to these latitudes.

As can be seen in Figure 1, the predominant climates of the tropics are dry ones: the hot deserts, semi-desert scrub and expansive savannahs. These lie towards the periphery of the tropical zone, in regions where anticyclonic subsidence predominates and rainfall is either seasonal or ephemeral. The hot deserts are noted for a high diurnal range of temperature. A maximum of more than 40°C in places by mid afternoon may fall as much as 30 degC by morning. This range is solely due to the effects of a dry atmosphere with little or no cloud and such a low vapour pressure that diurnal temperatures can vary greatly.

Over continental areas, the periphery of the humid zone is dominated by monsoon wind regimes. These bring wet humid weather in summer and predominantly dry weather in winter. Characteristically, there is a seasonal reversal of wind at low levels. In the northern summer, southeasterly winds cross the Equator and recurve to become south-westerlies (north-easterlies become north-westerlies in the Southern Hemisphere), bringing moist oceanic air across the Equator into areas that are under the influence of dry continental easterly winds during the winter. West Africa, southern Asia and northern Australia all experience these monsoon reversals. Although still seasonal, the situation is more complex over the Amazon basin and Caribbean, where moist westerlies cannot become established, due to the Andes-Sierra Madre mountain barriers. Thus the motion of convection is dependent on more complex changes in the atmospheric circulation and the influence of the Caribbean Sea, which warms and cools more than the neighbouring Atlantic Ocean.

Over the warmest ocean areas, the tropics are characterized by the development of tropical revolving storms, associated with strong lower-tropospheric winds and (perhaps more importantly) heavy rainfall.

Although a relatively narrow zone¶ (no more than about 1500 km wide, except in South-East Asia), it is the humid equatorial zone that many associate with the tropics. In this zone, rainfall can be relied upon year round, as the ITCZ is never far away. Maxima are generally limited to the mid 30s

* Wind speeds in this series are quoted in metres per second. 1 ms^{-1} = 2 knots.

¶ The ITCZ is sometimes seen to split into two branches, each of which may be several hundred kilometres across. This is notable in the Indian Ocean and the western Pacific.

Celsius[‡] over land by the high water-vapour content of the air and minima are similarly restricted by cloudiness or the overnight formation of dew, mist or fog. Over the sea, temperature changes little day by day. Thus, throughout the tropics, the diurnal range of temperature is relatively small. Nevertheless, this climate is uncomfortable for most humans, who find it difficult to lose excess body heat in these conditions. Indeed, this energy-sapping weather is used as a test by élite British army and Royal Marines regiments, which carry out part of their training in Brunei.

Where seasonality is the main effect on rainfall, savannahs predominate. In these areas of extensive grassland, annual evaporation exceeds precipitation and trees grow only in stunted groves. These areas are home to relatively large populations in some parts of the world and agriculture is critically dependent on the summer rains (both locally and to re-charge river flows), so any reduction or failure of seasonal rainfall often causes notable famines, especially in recent years. Some areas that have some seasonality, but a predominantly maritime climate, such as the northern Caribbean, have an intermediate climate with extensive forest, as well as grassland. Elsewhere, steppe surrounds the arid deserts. Here few plants can grow, but there is sufficient rainfall or run-off to support agriculture and moderate-sized populations. Included in this climatic zone are the highlands of much of Arabia.

‡ The 'rule' explaining the limitation of temperature assumes a boundary layer near saturation. This is comparatively rare, even in the humid zone. The humidity mixing ratio (h.m.r.) locally reaches 20 g kg^{-1} in the humid zone, limiting temperatures. However, the May pre-monsoon season, for instance, brings humid air across India with an h.m.r. closer to 16 g kg^{-1}, allowing temperatures to rise to around 40°C under clear skies. In either case, it feels very close!

However, within each climatic zone there are important variations, due to orography, latitude and longitude. Some of the world's hot deserts receive most of their (meagre) rainfall in summer, others in winter – here the variation is mainly by latitude and altitude, with equatorward regions having a summer rainfall peak. Examination of the mean annual rainfall in the tropics reveals that there are significant differences on a broad scale within climatic zones, as shown in Figure 4. In general, the desert areas generally see less than 200 mm yr^{-1}, although Australia's dry interior is defined by annual rainfall less than 600 mm. The monsoon zones see between about 1000 and 2500 mm yr^{-1}, occurring in summer. The semi-deserts and oceanic areas under the influence of the sub-tropical anticyclones have a total rainfall between 200 and about 1000 mm yr^{-1}. Areas under the influence of upper-tropospheric troughs (see Parts 2 and 3) see about 1000 mm yr^{-1} and the ITCZ experiences totals above 1000 mm yr^{-1} with some areas seeing more than 3000 mm yr^{-1}. Orographic effects add further local detail.

High ground has two main effects: it lowers the mean temperature (although nights may be less cool, if the mountains are modest and the observation is not in a valley) and increases the rainfall, increasing the likelihood of precipitation by 'forced' convection or convergence. For instance, copious rainfall is generated by the Ethiopian Highlands and India's Western Ghats. These highlands have a relatively equable climate within the almost universally hot tropical zone and their climate is markedly different from that of their surroundings.

Along the western margins of Africa and the Americas, there are cool currents as a result of the upwelling of cool deep water under the influence of the Trade Winds. The effect is most marked along the south-west coast of South America, as well as the north-western and south-western coasts of Africa. This brings a surprisingly cool and usually dry climate north to only a few degrees south of the Equator, although this climatic zone, comparatively equable for the population, is narrow, extending only a few tens of kilometres inland. Populations are comparatively high in these zones, largely due to the excellent fishing available, due to the upwelling of plankton-rich deep waters.

Heating and cooling have a significant effect on air pressure throughout the tropics, the pressure falling in response to diurnal heating and rising in response to nocturnal cooling. As pressure rarely changes due to synoptic-scale weather systems in the tropics, the diurnal changes are significant, as shown in Table 1.

Clouds and fog in the tropics

Most *Weather* readers will be familiar with the clouds observed in the middle latitudes and their typical range of heights (Pouncy, 2003). The clouds of the tropics are the same types as those seen in the rest of the world, although convective clouds are predominant, whilst altostratus and nimbostratus are relatively rare. However, the increase in the depth of the troposphere means that the range of height at which each species is seen is greater than in the middle latitudes. The World Meteorological Organization (WMO) recommends the range of cloud height given in Table 2 (WMO, 1956).

All stratocumulus clouds[†] in the tropics are at a temperature above 0°C and it is altocumulus and altostratus (as well as occasional nimbostratus) that occupy the level of

† This assumes that the cloud height is not relative to the observer. Clearly there are layer clouds at heights less than 2000 m above high ground in the tropics that are at or below 0°C. Typically, this would be the case for an observation made at an altitude above 2500–4000 m in the tropics.

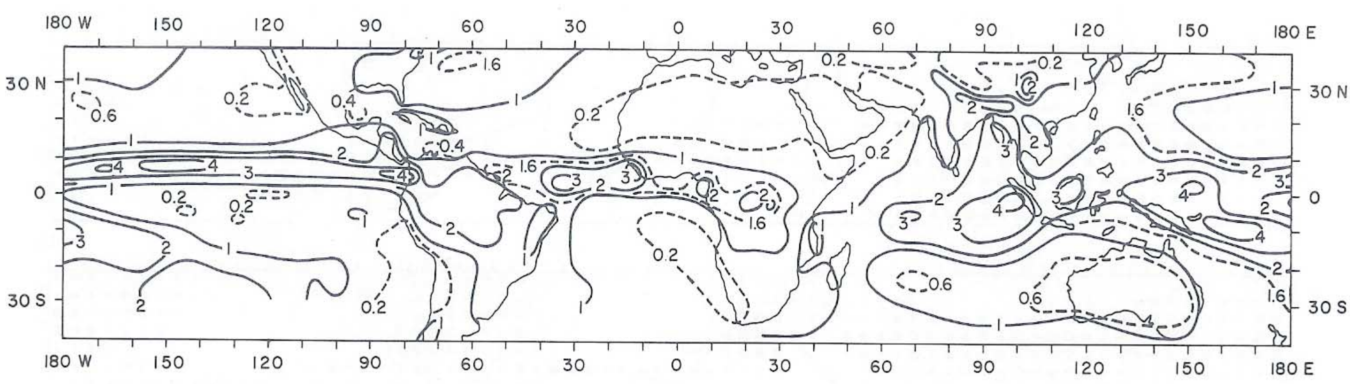

Figure 4. Mean annual rainfall (m) in the tropics (from Hastenrath, 1991, Figure 6.1:7). Local orographic effects are not evident at this scale, although a rainfall maximum in excess of 2000 mm yr^{-1} is shown over the southern Himalaya.

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Table 1

Typical pressure changes due to the heating and cooling of the atmosphere in the tropics.

Correction to the observed pressure necessary to allow for diurnal variation			Average component of the 3-hourly barometric change due to the diurnal variation		
Local time	0°–10°N or S	10°–20°N or S	Local time	0°–10°N or S	10°–20°N or S
	HPa	*hPa*		*HPa*	*hPa*
0000	−0.6	−0.5	0000-0300	+1.1	+1.0
0100	−0.1	−0.1	0100-0400	+0.9	+0.9
0200	+0.3	+0.3	0200-0500	+0.3	+0.3
0300	+0.7	+0.7	0300-0600	−0.5	−0.5
0400	+0.8	+0.8	0400-0700	−1.2	−1.1
0500	+0.6	+0.6	0500-0800	−1.5	−1.4
0600	+0.2	+0.2	0600-0900	−1.5	−1.3
0700	−0.4	−0.3	0700-1000	−1.0	−0.9
0800	−0.9	−0.8	0800-1100	−0.2	−0.4
0900	−1.3	−1.1	0900-1200	+0.7	+0.6
1000	−1.4	−1.2	1000-1300	+1.5	+1.3
1100	−1.1	−1.0	1100-1400	+1.8	+1.7
1200	−0.6	−0.5	1200-1500	+1.9	+1.6
1300	+0.1	+0.1	1300-1600	+1.4	+1.2
1400	+0.7	+0.7	1400-1700	+0.7	+0.5
1500	+1.3	+1.1	1500-1800	−0.3	−0.2
1600	+1.5	+1.3	1600-1900	−1.0	−1.0
1700	+1.4	+1.2	1700-2000	−1.5	−1.4
1800	+1.0	+0.9	1800-2100	−1.6	−1.5
1900	+0.5	+0.3	1900-2200	−1.4	−1.1
2000	−0.1	−0.2	2000-2300	−0.8	−0.6
2100	−0.6	−0.6	2100-2400	0.0	+0.1
2200	−0.9	−0.8	2200-0100	+0.8	+0.7
2300	−0.9	−0.8	2300-0200	+1.1	+1.0

Table 2

The variation of range of cloud-base heights in tropical, middle-latitude and polar air masses.

Étage	*Polar regions*	*Middle latitudes*	*Tropical regions*
High (Ci, Cs, Cc)	3000–8000 m	5000–13000 m	6000–18000 m
Middle (Ac, As, Ns*)	2000–4000 m	2000–7000 m	2000–8000 m
Low (St, Sc, Cu, Cb)	0–2000 m	0–2000 m	0–3000 m

* Nimbostratus usually has a base that extends down into the low-cloud étage and a top that is frequently well into the high-cloud étage.

transition to supercooled water. The tops of the deepest medium-level clouds may contain a significant amount of ice as the temperature of the cloud falls below about −20°C. This temperature occurs at an altitude between 8 km and, locally, 9 km. Clouds formed predominantly of ice are not found below 8 km in the tropics.

Cumulonimbus clouds always have tops extending into the middle levels and frequently into the high-cloud étage. This is the usual state in the tropics, where cumulonimbus clouds rarely have tops below 12 km. In hot-desert regions, where there is sufficient moisture, cumulus or, if there is sufficient moisture at high levels, cumulonimbus clouds, may form around the time of maximum temperature.

Precipitation from cumulonimbus clouds is usually heavy. These clouds, resulting from air with a high humidity mixing ratio ($r \cup 20$ g kg^{-1} in the ITCZ), can produce large amounts of precipitable water and are the source of thunderstorms of the ITCZ. Thunderstorms are frequently seen in association with the deep convection of the tropics. However, they are more common near the poleward edge of the ITCZ, where instability and convective available potential energy (CAPE) values (defined in Galvin *et al.*, 1995) are larger than nearer to the Equator, especially over the continents (see Figure 5). Despite the presence of cumulonimbus, which may produce hail, much of the tropics is simply too warm for hail to reach the ground, although it does occur in places where wind shear and vorticity are conducive. Waterspouts and tornadoes may also form, but only where there is sufficient vorticity.

From cumulus clouds, precipitation is generally slight. Nevertheless, this rainfall (or, over high mountains, snowfall) can form the majority of annual precipitation.

Stratus and fog are also characteristic of parts of the tropics. Stratus with hill fog is frequently seen in the early morning over tropical woodland, in particular the tropical rainforest, following overnight cooling. Here, transpiration from the trees in a very moist environment assists its formation.

Fog and stratus are also features of cooler parts of the tropical oceans and are frequently seen along the east-facing coasts of southern Africa and South America. In the former case, much of the meagre precipitation available for plants to grow in the Namib Desert is deposited from wet fogs.

Early-morning fog is also relatively common on the coasts of the Arabian Gulf, where sea breezes frequently bring moist air inland during the afternoon, humidifying the air and raising its fog point. Extreme summer cases (observed in Doha and Bahrain), when sea-surface temperatures in the Gulf may reach 35°C, may bring fog with visibility less than 100 m around dawn with air temperatures and dew points above 30°C (Andy Dexter, Richard Young personal communications)! These fogs feel very moist, almost suffocating, making it difficult for humans and animals to lose excess heat. As such, they can be dangerous to life in a very different way from fogs in the middle latitudes.

Layer clouds are usually the product of convection and are often relatively thick, as a result. Most of these clouds of convective origin are of genera altocumulus or stratocumulus. However, there are some exceptions and altostratus or nimbostratus sometimes form, usually in association with cyclonic disturbances. These will be discussed later in the series. Hours of steady precipitation may fall from these deep layer clouds.

Altocumulus castellanus (or floccus) clouds are also characteristic of tropical air. Principally, they form in response to high-level cooling. This occurs in two ways. First, there is the dynamical cooling associated with advection, often ahead of upper troughs. This is usually evident above the 700-hPa level. Second, there is the long-wave cooling of the atmosphere during the evening, causing a decrease in stability at the cloud tops. Castellanus development may prolong the life of cumulonimbus through the night. In some cases, the destabilization is associated with instability at lower levels, which feeds into the unstable medium-level clouds.

Any unstable cloud whose top becomes glaciated is reclassified as cumulonimbus. Thus the base of these cumulonimbus clouds, originally altocumulus castellanus, may be at 4000 to 5000 m or more. Most

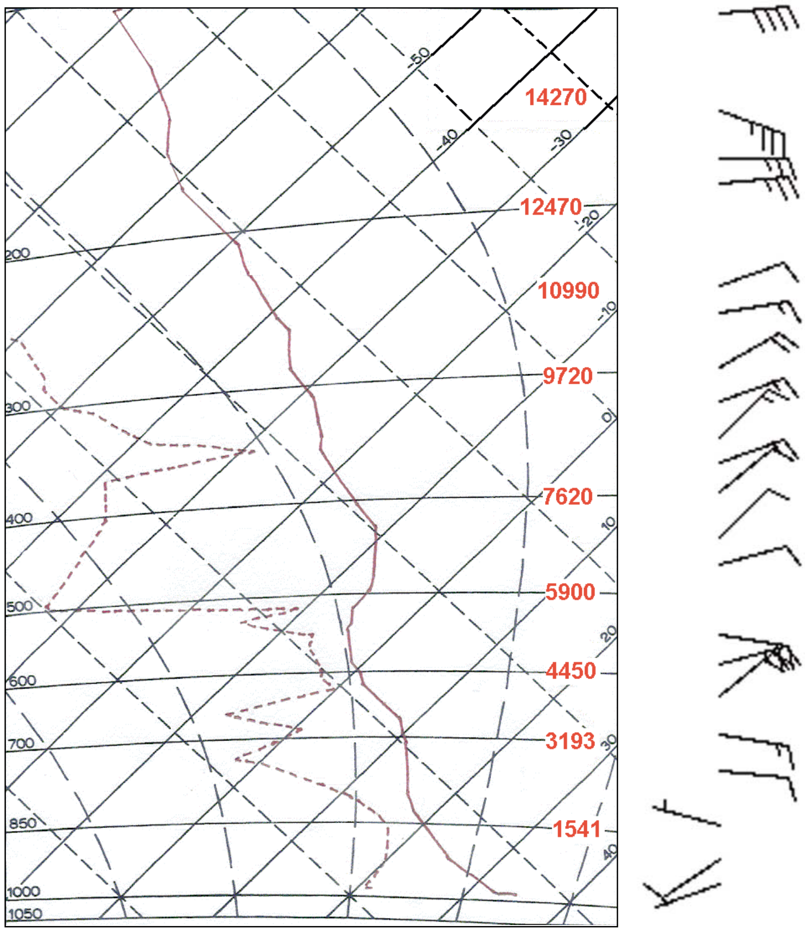

Figure 5. Radiosonde profile for Niamey, Niger, in Africa's Sahel at 1200 UTC on 30 July 2006. This ascent is absolutely unstable and has a convective cloud base near 850 hPa, in contrast to ascents typical of the ITCZ. It shows a typically high level of CAPE (1273 J kg^{-1}) and θ_w falling with height.

Box 1. The effect of latitude on geostrophic balance

Over much of the globe, wind speed and direction are the result of a balance of forces, the most significant of which are the pressure-gradient force and the Coriolis force, which are in equilibrium where the flow is straight and parallel to the Equator (Persson, 2000). However, this relationship breaks down close to the Equator, where the Coriolis force becomes very small, since the air is subject to very little rotation around the Earth's axis.

Nevertheless, winds remain in reasonable balance until they reach the latitudes of about 6° and even then momentum generally carries the wind in the direction it was moving when in near-geostrophic balance. Bigger problems for the analyst include the distance between isobars, which become very large in the tropics, such that barometric errors (and those of correction to sea level) make standard analysis, based on mean sea-level pressure (or geopotential height), almost impossible. As a result, most analysis in the tropics uses streamlines.

If we define the tropics as the zone between the sub-tropical discontinuities, the tropical atmosphere is barotropic, so horizontal temperature differences are small. Furthermore, the apparent progress of the sun from east to west across the sky causes a pressure disturbance as the air is heated and cooled by incoming and outgoing radiation. The wavelength of this disturbance is 12 hours and its amplitude is approximately 2 hPa at the surface (Table 1); the amplitude is somewhat larger than the usual change of pressure in the atmosphere.

Over land areas, the use of mean sea-level pressure is also inappropriate since the correction of observed pressure to sea level causes errors. Thus, where geostrophic balance needs to be measured, the height of the 925 hPa or 850 hPa surface should be used.

thunderstorms that develop over the hot deserts of the tropics have such high bases. Precipitation from these high-based cumulonimbus clouds usually evaporates before reaching the ground and strong downdraughts are the result.

Population in the tropics

On average, population density is low in the tropics: typically around 10 per km^2. However, in India, Bangladesh, southern China and parts of South-East Asia, the population density is much higher, reaching 100 per km^2 or more.

Away from rivers, the hot-desert environment is not able to support large populations and tropical rain forest is a dark foreboding environment with trees that are difficult to clear, unless heavy machinery is available. Thus populations are found mainly along coasts or rivers, where transport has been available for centuries or millennia. It is near the mouths of the larger rivers that the largest cities are found.

The savannahs are moderately populated, the land easy to clear for agriculture and road building. However, in some areas the population density is close to the ability of the land to support it. In Africa and parts of South America, this presents a problem, since populations have grown rapidly during the late twentieth century and the rainfall is not sufficiently reliable in these areas for there to be confidence that populations can survive without a major risk of drought and famine.

The degree of urbanization, by contrast, varies considerably from continent to continent. In South America, it is generally above 50%, the proportion in Venezuela above 90% and in Australasia 80%. But in much of Asia and Africa it is below 45%, Ethiopia, Uganda, Malawi and Nepal having fewer than 15% of their populations living in towns. This stresses the relative importance of agriculture in these countries, even those that are rapidly industrializing, such as India and China. However, there are significant variations from country to country and within countries.

El Niño and the Southern Oscillation

The El Niño phenomenon is generally accepted to have a profound underlying effect in the tropics, changing the broad-scale circulation patterns as pressure changes across the Pacific to form the Southern Oscillation. The Southern Oscillation is a relative change of pressure in the eastern tropical Pacific basin (usually measured at Tahiti) compared with that in the west (measured at Darwin). El Niño is a warming of the eastern equatorial Pacific Ocean and causes mean sea-level pressure to fall over the eastern Pacific, whereas over the western Pacific, pressure rises.

The consequence of the changes is to increase precipitation in the eastern Pacific (and over the Atlantic and western Indian Ocean basins), while decreasing it across South-East Asia and Australia (as well as, perhaps, east Africa).

Relatively strong westerlies are associated with the development of the El-Niño anomalies in the eastern Pacific (Verbickas, 1998; Fedorov, 2002) and their formation mechanism seems to be directly linked with the El-Niño phenomenon. In view of the mass transport associated with such a change in wind velocity, accompanied by the vast amount of water vapour that can be carried by such tropical winds, it is easy to see how El Niño has global consequences.

La Niña is an amplification of the westward flow of the Trade Winds and associated increased warmth of the western Pacific. It is measured as an increase from the normal pressure difference Tahiti–Darwin. Its consequences are an increase in west Pacific tropical storm activity and copious additional rainfall across South-East Asia and north-eastern Australia.

A review of El Niño and its global consequences appeared as a special issue of *Weather* in 1998 (Vol. **53**, No. 9), so no more will be said here.

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A particularly threatening advancing squall line, as seen from Stratfield Mortimer, Berkshire, looking south-west at 1812 UTC on 19 August 2006. The surface dew point temperature fell rapidly, 2.5 degC in 7 minutes, as the leading edge of the squall line passed overhead; this was followed by two hours of showery rain starting a few minutes after the photograph was taken. (© Stephen Burt.)