Chapter 3

Himalayan weather systems & temperature controls

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1 Introduction

The atmospheric circulation systems affecting the Karakoram, Hindu Kush and western Himalayas are different from those affecting the central and eastern Himalayas. In winter, the western Himalayas are affected by the 'Westerly Disturbances' (WDs) which derive from the Atlantic, Mediterranean and Black Sea systems[1]. These storm tracks will only invade the central Himalayas every few years.[2] This is essentially a different system from the central and eastern Himalayas, whose winter weather is derived from cold dry north easterlies and whose summer weather derives from the Arabian Sea to the west. The summer monsoon[3] arrives in India as a

[1] Hatwar, H. *et al.* 2005.
[2] Seko, K. & Takahashi, S. 1991.
[3] The word *monsoon* derives form an Arabic word meaning season. Thomas, D. & Goudie, A. 2000.

south-westerly, but turns north-west over the Indo-Gangetic Plain (IGP) entering Nepal as a south-easterly.

3.2 Weather patterns affecting the Karakoram, Hindu Kush and western Himalayas

The Karakoram hydrological and climatic regimes are somewhat different to much of the rest of the Himalayas. The majority of the precipitation arrives in winter and spring rather than summer, dominated by systems from the west that originate from the Mediterranean and Atlantic. While the great floodplains of Pakistan are hit by the southerly and south-westerly summer monsoons, these have usually lost much of their moisture by the time they reach the Upper Indus Basin (UIB). Although they contribute to the system, over 80% of the upper Indus consists of snow and glacier melt. Thus, the monsoon is not usually the dominant force in summer precipitation for the UIB, leaving it vulnerable to climatic variability.[4] Despite this there has been a slight increase in summer rainfall, particularly in June.[5] The majority of stream flow in the mid hills results from summer runoff derived from the ephemeral nature of previous seasons' snow pack. Seasonal increases in temperature at this altitude tend to lead to sublimation or evaporation and actually reduce stream flow (depending upon ambient atmospheric moisture). At higher altitudes this situation is reversed: snow and ice melt becomes the chief water source. Each 1°C increase produces a 16 –17% increase in runoff.[6]

The Indian (and therefore the central Himalayan) summer monsoon seldom reaches over into the Karakoram from the east. When, on occasions with a multi-decadal return period, it does penetrate significantly into the Hindu Kush, it interacts dramatically with the features of the westerly systems. This was evidenced by the July – August floods of 2010 in Pakistan.[7]

In much of the rest of the Himalayas, increases in mean winter minimum temperatures are proportionally greater than increases in mean maximum temperatures, leading to a reduction in diurnal temperature ranges (DTR) and increases in mean annual temperature (MAT). By contrast, DTRs in the Karakoram are increasing, particularly in autumn. MATs are shown to be falling as a result of stronger summer cooling than winter warming. The fact that some glaciers in the Karakorum are showing reverse trends to other areas of the Himalayas attracts the attention of glaciologists and climate change sceptic alike. Fowler and Archer have demonstrated that since the beginning of the 1960s winter mean temperature has increased, but summer mean temperature has dropped as much as 1°C, causing a 20% decline in runoff for some rivers.[8] This anomaly is probably associated with recent lowering in elevation of the moisture bearing westerly jet stream.[9]

Increases in winter-spring precipitation in the UIB give rise to snow accumulation rates of 1.5 to 2m^{-a} at 5500m.[10] Spring mean minimum temperatures and snow cover are often cited as being

[4] Archer, D. & Fowler, H. 2004.
[5] The source of this precipitation has yet to be attributed to westerlies or monsoonal incursions. Personal communication from Nathan Forsythe, Newcastle University, 8 Feb. 2011.
[6] Fowler, H. & Archer, D. 2006.
[7] Personal communication from Nathan Forsythe, Newcastle University, 8 Feb. 2011.
[8] Fowler, H. & Archer, D. 2006.
[9] Scherler, D. *et al*. 2011 citing work by Archer & Caldeira
[10] Fowler, H. & Archer, D. 2005.

crucial to the relative 'health' of a glacier. It is likely therefore that the cooling spring mean temperature observed at some (though not all) stations is contributing to the building of glaciers and glacier surges.[11]

3. 3 The Indian summer monsoon

The Indian summer monsoon is among the most impressive weather systems on earth. Yet despite considerable progress in modelling and understanding over the last decade, uncertainties remain both in aspects relating to natural cause and effect on the one hand and of anthropogenic forcing on the other. In Nepal such uncertainties are accentuated at all scales by topography and it is particularly important to keep this in mind when reading paragraphs 3.3 and 3.4.

3.3.1 Propagation of the summer monsoon: At its simplest, the Himalayan summer monsoon is a function of the thermal contrast between a warming land mass and a relatively cooler ocean. Much of the summer monsoon originates in the Arabian Sea as a south-easterly to the east of Madagascar and is deflected at almost 180^{o} by northern Madagascar and the Ethiopian Highlands. The Coriolis force also plays an important part in this eastward deflection.

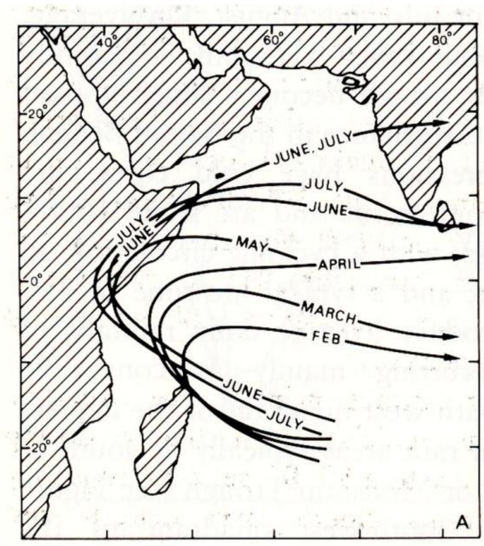


Figure 3.1: The mean monthly positions of the low level Somali jet stream over the Indian Ocean.[12]

This airflow becomes extremely powerful and can run at up to 45m s^{-1} (100mph) at levels of 1000 – 1500m,[13] with a force of at 850hPa.[14] Over the period between February and July the westerly flow is pulled north from about 5^{o}S in winter to about 20^{o}N in summer by the northward excursion of the easterly airflow of the Inter Tropical Convergence Zone (ITCZ) (see Fig 3.2.)

[11] Glacier surging is a phenomenon which is now thought not to be related to temperature fluctuations but rather driven by the build up of a reservoir of ice in the accumulation zone that will discharge unsustainably until exhausted and then begin to re-accumulate. Geology, gradient and relationship of length to width are also important as factors (Benn & Evans 1998 pp. 169-171).

[12] Barry, R. & Chorley, R. 1992, p. 250.

[13] Barry, R. & Chorley, R. 1992.

[14] Goswami, B. 2005. hPa (hectoPascal) is a unit of pressure =100 Pascals = 1 Newton/m^2 is equivalent to 0.1% of atmosphere.

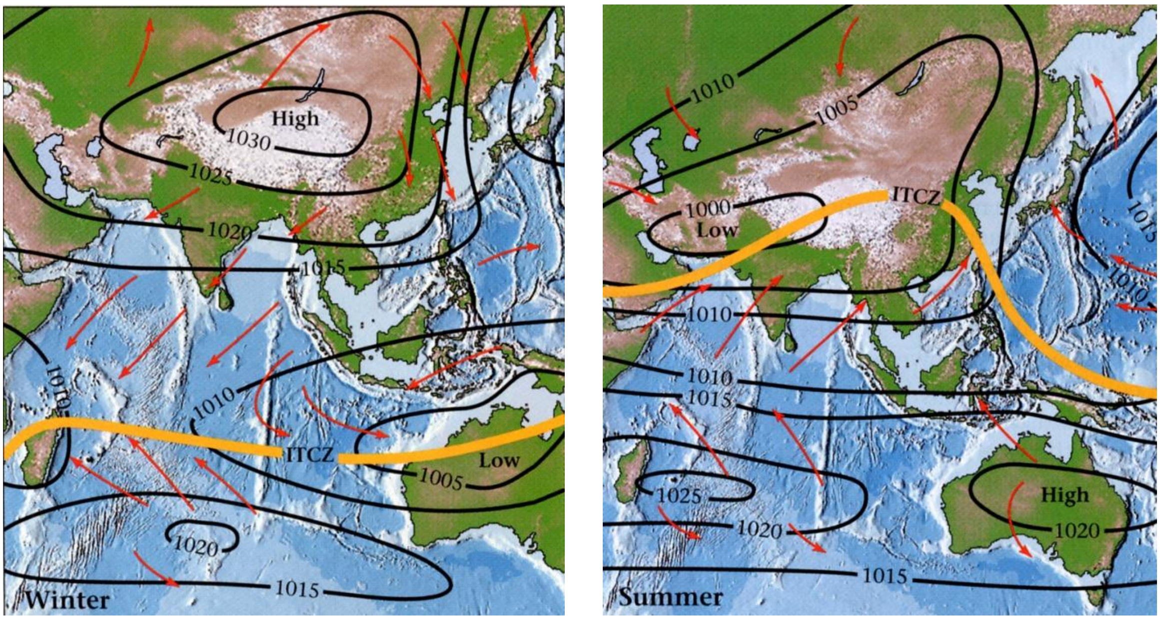


Figure 3.2: The orange line shows the excursion of the ITCZ from winter (south of the equator) to summer, (trans Himalaya to the northern TP).[15]

3.3.2 Predicting the summer monsoon

Gauging the onset of the monsoon has probably been a problem since mankind first arrived in India. The summer monsoon brings 80% of the subcontinent's precipitation and its arrival is of critical importance to over a billion people. Present day forecasting builds in a large number of variables the results of which frequently remain uncertain.[16]

Burroughs reports that the monsoon was less predictable pre-1920 than it is now, for example causing the massive famine in India in 1878, after which the British Government called upon the H.F. Blandford of the meteorological department to determine the means of making forecasts. He concluded that heavy Himalayan snow into the spring would minimise the summer monsoon but that weak snow cover would maximise it. While this theory went out of fashion for a period in the 1950s, it is now recognised as an observation of considerable significance and forms part of the accepted theory by which the warming Tibetan Plateau (TP) draws up moist air into the north of the subcontinent. The reduction in seasonal snow cover and earlier snowmelt observed on the TP in recent decades will be important variables in any aspect of this study.

But it is not just the snowfall on the TP that is significant. Barnett and colleagues[17] have shown the importance of the snow cover across the whole of the Eurasian continent, with particular reference to the depth of snow. Using data from the very different Asian snow-cover extents in 1978 and 1970 (13-year maximum and minimum respectively) they showed that a 2cm cover had only local atmospheric significance. However modelling heat requirements to melt variables of observed increased snow depths produced significant

[15] Marshak, S. 2001 p. 640.
[16] Burroughs, W. 1999 p.138.
[17] Barnett, T. et al. 1989

correlations to observed atmospheric patterns. This is because of (a) enhanced albedo and (b) because the soil will remain cold and wet for up to another month, once a deeper snow pack has melted, reducing reflected heat. If this period runs past the summer solstice, the amount of radiation available for evaporation begins to decline. Further, the potential of the arid atmosphere will delay precipitations and thus the release of latent heat.

A different approach to the complexity and interconnectedness of the system has been set out by Goswami,[18] who describes a number of aspects of monsoon variability. Perhaps the most important is the delay of the monsoon by a 'bogus' monsoon which appears to present an early onset that subsequently fades, with catastrophic consequences for agriculture. In this case an early first phase northward migration of the ITCZ may be determined by Intra Seasonal Oscillations (ISOs) caused by sea surface temperature anomalies in the south tropical Indian Ocean and western equatorial Pacific.

The longer oscillations tend to have a wider geographical range than the shorter ones. But because of their quasi periodic nature, ISOs remain difficult to predict. ISOs of between 10 and 90 days have been identified as the drivers for Intra Seasonal Variations (ISVs), longer spells of wet and dry conditions of 2 – 3 weeks 'active' and 'break' periods. Breaks can reduce agricultural yield. Models are being presently being developed which have the potential to predict ISVs up to three weeks in advance.

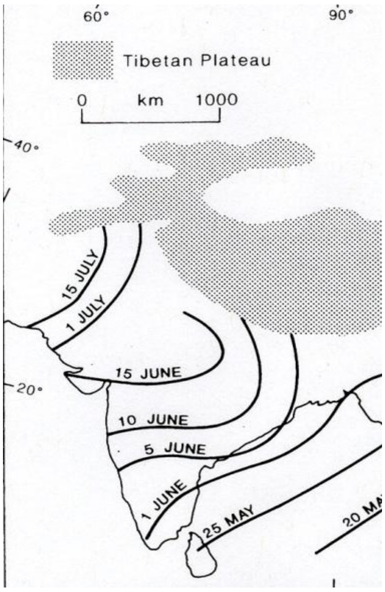
The break condition is associated with a decrease in cyclonic vorticity and an increase in surface pressure over the monsoon trough in conjunction with a weaker lower level jet stream. Weakening of the Tibetan anticyclone in the upper atmosphere is also though to engender breaks.

The influence of El Niño on the summer monsoon is discussed under 3.4.1.2

3.4 The summer monsoon and the Himalayas: Introduction

The monsoon arrives in India as a south-westerly. Yet it hits the central part of the Himalayas predominantly as a south-easterly having drawn up further moisture from the Bay of Bengal. On reaching the Himalayan foot hills, orographic lifting will lead to cloud formation and intense rainfall.[19] However its influence on river systems gradually diminishes from east to west. As the monsoon is the mainstay of life in the Himalayas, it is worth understanding how this air flow is deflected and why it enters much of Nepal from the south-east.

[18] This and the following two paragraphs are derived from Goswami, 2005. (There may however be slight confusion in the way he cites figure 2.1 in his text.)
[19] Marshak, S. 2001.



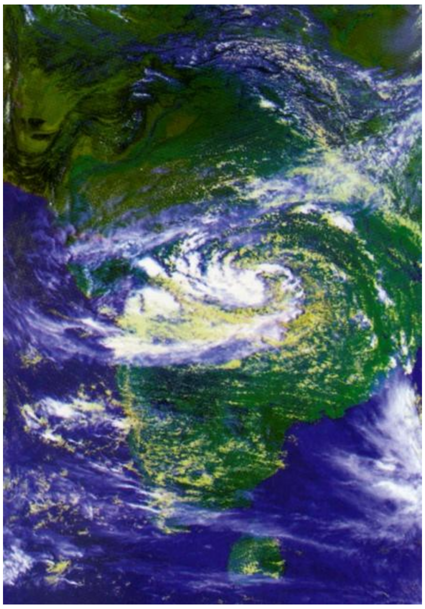


Figure 3.3: The onset of the summer monsoon

Monsoon arrival dates.[20]

NOAA-14 monsoon image of the subcontinent showing depression.[21]

Factors driving the northwards swing of the monsoon may be the warming of the TP, which, because of its elevation to about 5km, has been thought to exert a stronger influence on the northward flow of the summer monsoon than would be exerted by low lying terrain. The situation may be more complex however, since the highest upper tropospheric temperatures have been found over northern India rather than above the TP and 'about 1,000km west of the peak precipitation'. Effectively the Himalayas insulate the moist tropical air of the monsoon from the dry colder (katabatic) air from the TP.[22]

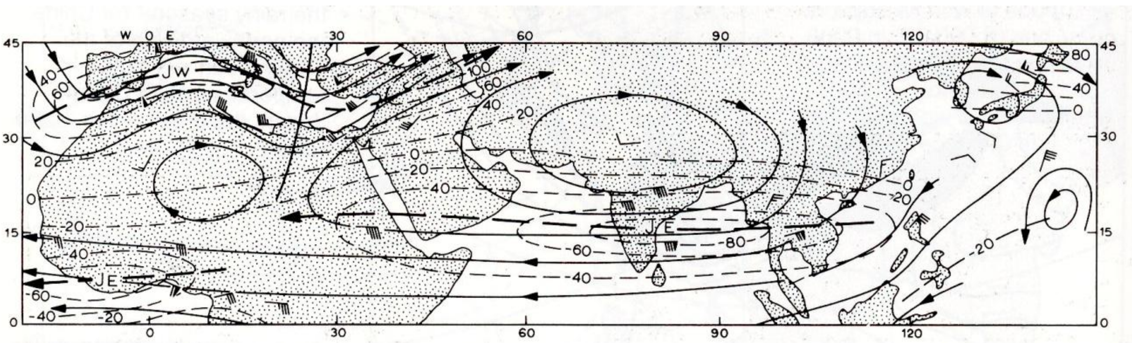


Figure 3.4: Higher summer circulation.[23]

One of the main driving forces for this change of direction is thought to come from strong, high easterly airflows overlaying the monsoonal flows which work in a clockwork circulation. This also appears to influence differences in summer rainfall between the east and the west of the Himalayas. In the east, the south-westerly airflows are less affected. 'In the northwest a thin wedge of monsoon air is overlain by subsiding continental air. The inversion prevents

[20] Barry, R. & Chorley, R. 1992, p.289.
[21] Burroughs, W. 1999, p.139.
[22] Boos, W. & Kuang, Z. 2010.
[23] Barry, R. & Chorley, R. 1992 p. 284.

convection and in consequence no rain falls in the summer months in the arid north-west of the subcontinent.'[24]

Whether the influence of this factor is strong enough to cause the slightly lower summer rainfall in the north west of Nepal is unclear. More probably, the strength of the summer monsoon will become dissipated as it reaches Nepal's north west. It gives the central region (Pokhara) the highest mean annual rainfall at about 5m, the vast majority of which falls during the monsoon. (See Figure 3.5 below)

Part of the summer monsoon arrives as a southerly across the Bay of Bengal from which it can gather additional moisture. This also contributes to eastern Nepal (either side of Latitude 28°N) receiving higher annual rainfall than the west.[25]

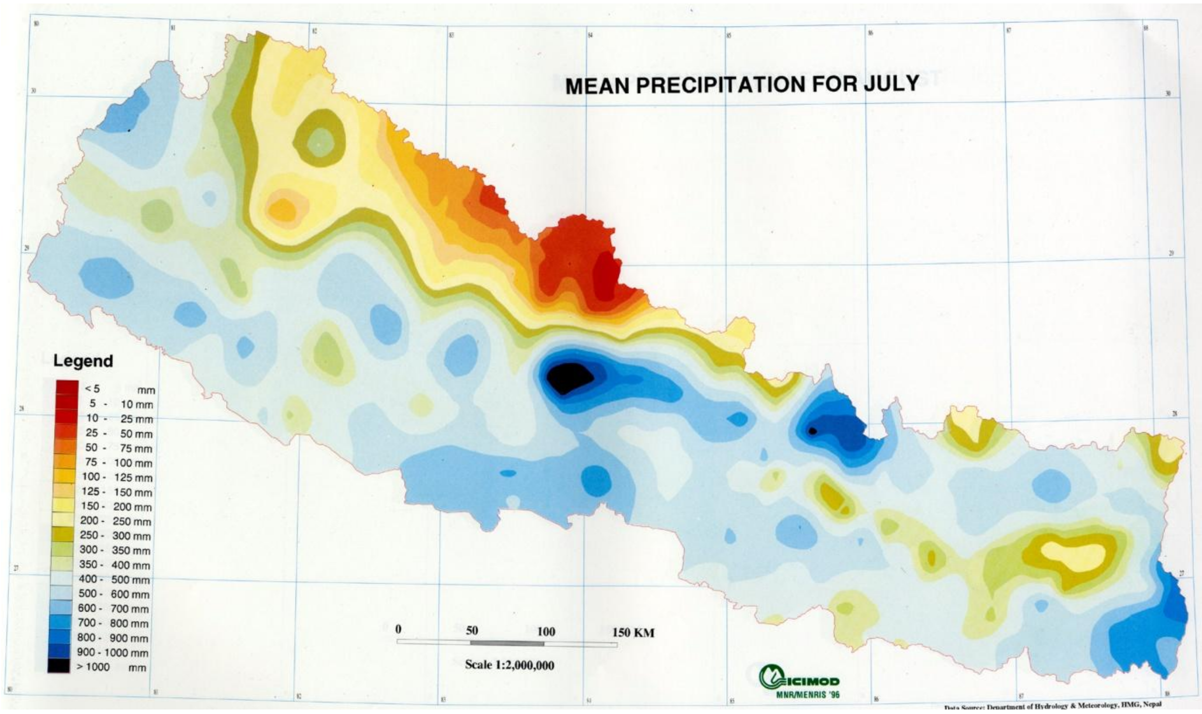


Figure 3.5: Mean Precipitation for Nepal in July.[26]

3.4.1 Factors potentially affecting Nepali summer monsoon variability

Some of the factors which have the potential to affect the Nepali summer monsoon have already been discussed in terms of the Indian monsoon under 3.3. However there are other factors which affect the monsoon as it pertains to northern India and the Himalayas. Two such factors are discussed below.

[24] Barry, R. & Chorley, R. 1992 p. 247.
[25] Shrestha, A. & Kostaschuk, R. 2004.
[26] Chalise *et al.* 2005 p. 101.

3.4.1.1 Aerosol forcing and the summer monsoon[27]

Aerosols, tiny particles suspended in the atmosphere, have both natural and anthropogenic origins. The two main anthropogenically produced aerosols that concern us here are black carbon and (white) sulphates. Broadly speaking the first is absorptive and the second reflective. Both have short atmospheric residence periods of 'a week or two'.[28] They usually have a strong regional signature.[29] At its simplest, the aerosol number density (referred to as aerial optical density, AOD) that is being presently experienced has the capacity to reduce incoming shortwave solar radiation by up to 10%,[30] cooling the atmosphere below and warming the atmosphere above, both by absorption and enhanced cloud albedo.

Naturally occurring dust and anthropogenic black carbon both absorb shortwave (incoming) radiation, reducing surface temperatures but warming the overlying atmosphere. Black carbon also has the capacity to amplify the effect of some other anthropogenic aerosols by a factor of three.[31] It has a high capacity to intercept UV and visible wavelength radiation, causing a reduction in surface radiation and reducing evaporation. The effect can be considerable, with surface cooling in the order of 10% and doubling rates of higher atmospheric warming. Sulphates (non-absorbing aerosols) scatter shortwave radiation, causing cooling at the surface, but relatively little atmospheric warming.[32]

Because aerosols reduce sea surface temperature (SST) the monsoon's ability to absorb moisture from the Arabian Sea is thought to be affected. (This aspect of the aerosol process is referred to as the *direct effect*.) During the summer monsoon, aerosols from naturally occurring dust storms in the Horn of Africa and the Arabian Peninsular are brought in on westerlies, covering the IGP and reaching into the Himalayas.[33] These have the potential to combine with more locally derived natural aerosols (e.g. from the Thar Desert) and form a west-east gradient along the southern slopes of the Himalayas.[34] Combined with anthropogenic aerosols, AOD can peak at 0.6 during July.[35]

There are different opinions as to the anthropogenic fraction of such peaks[36] and their formation into 'atmospheric brown clouds' (ABCs). Industrial production of southern Asia has led to an estimated six fold increase in atmospheric black carbon between 1930 and 2000.[37] Such ABCs

[27] Excellent introductory information on aerosols is to be found on the NASA website (see Chapter 3 Bibliography).
[28] Ramanathan, V. *et al.* 2005.
[29] Prasad, A. *et al.* 2005.
[30] Lau, K.M. *et al.* 2008
[31] Ramanathan, V. & Crutzen, P. 2003.
[32] Lau, K. M. *et al*, 2006
[33] Prasad, A. *et al*. 2009.
[34] Shrestha, P. & Barros, A. 2010.
[35] Li, F. & Ramanathan, V. 2002. However this may vary on a regional scale. See Gautam, R. *et al.* 2009 who find highest AODs in north-west India during May.
[36] Satheesh, S. & Srinivasan, J. 2002 find that over India's adjacent oceans, the predominance of natural aerosols as a fraction of the whole only occurred between June and September. Gautam, R. *et al.* 2009 find along broadly similar lines for the north of India, although with higher anthropogenic weightings in the post monsoon period. Ramanathan, V. & Crutzen, P. 2003 suggest that 10-14% of ABCs is composed of black carbon.
[37] Ramanathan *et al.* 2005

can be extensive, covering 'most of the Arabian Sea, Bay of Bengal, the Northern Indian Ocean and the South Asian region and extend[ing] from about November to May and possibly longer.'[38] While ABCs have the potential to offset the lower atmosphere from the effect of GHGs by up to 50%, they effectively form a blanket to evaporation over the Arabian Sea. Ramanathan *et al.* suggest that if these trends continue, the Indian subcontinent may experience a doubling of drought frequency. They note that since the 1960s the Indian monsoon rainfall, most notably in July, has declined by something of the order of 5%, compared with the 1930 –1960 average, very much in line with their model.[39] Aerosols are thought to inhibit atmospheric convection which, when enhanced by stable conditions such as monsoon breaks, may create subsidence or inversion layers.[40]

The relationship of aerosols both to cloud formation and precipitation is complex and may not yet be well understood. Much will depend on the type of aerosol. Sulphates, which are hygroscopic, provide enhanced cloud condensation nuclei (CCN) and may promote precipitation, whereas black carbon and natural dust are hydrophobic and may suppress precipitation.[41] Some of the differences in aerosol 'behaviour' reported in the literature may be conditional upon aerosol composition although this is an aspect of the complexity which seldom seems to be discussed.

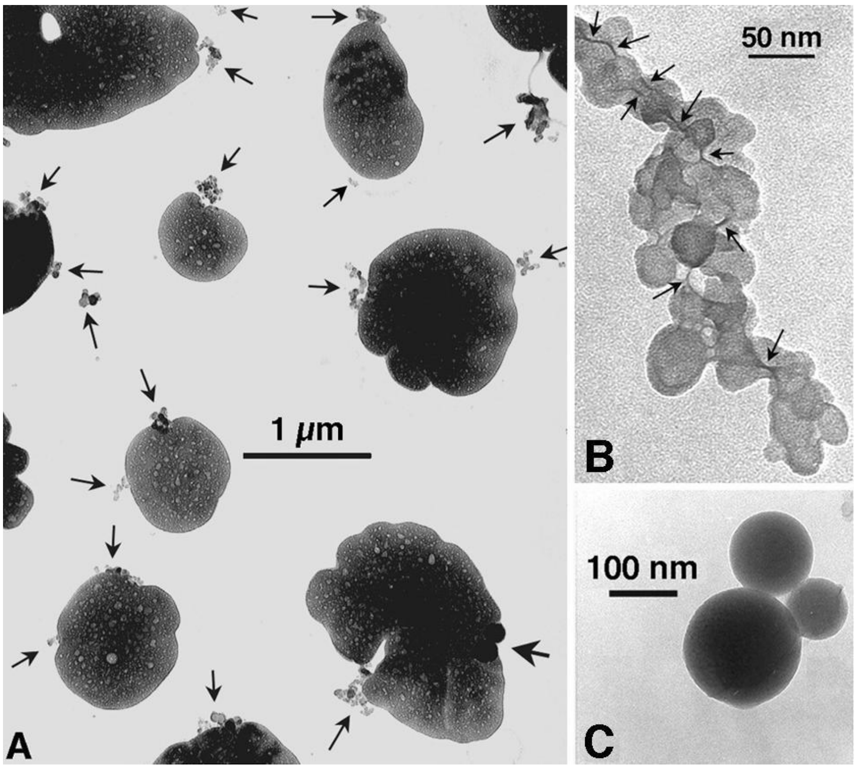


Figure 3. 6:[42]
A: 'White' sulphates (arrows pointing at black carbon chains)
B: Black carbon chains
C: "Fly ash" formed from coal burning
1μm= 1/1000 mm
1nm = 1/000,000 mm

Increasing AOD will provide more CCN which can negatively affect cloud effective radius[43] (referred to as the *indirect effect*) potentially inhibiting rainfall, particularly over the ocean. It is

[38] Ramanathan, V. & Crutzen, P. 2003, p. 4003.
[39] Ramanathan, V. *et al.* 2005.
[40] Lau, K.M. *et al*. 2008.
[41] Gautam, R. *et al.* 2009.
[42] NASA (a): Picture credit: Peter Buseck, Arizona State University, from an (undated) paper by Pósafi *et al*.

further thought that enhanced CCN number density reduces cloud size but increases the area of cloud distribution. These so called 'bright clouds' have high albedo but may be less prone to precipitation, unless in an already humid environment.[44] This appears to be particularly true of black carbon. The effects of aerosols may even be different over ocean and over land. For example over conurbations and immediately downwind of potential aerosol sources rainfall may be heavier because of aerosol nucleation resulting from stronger atmospheric circulation.[45]

> [In]…Central Nepal including the IGP…there is consistent spatio-temporal agreement among aerosol, cloudiness and rainfall variability both in pre-monsoon and monsoon seasons, and thus … the aerosol indirect effect is likely to be important in the region.[46]

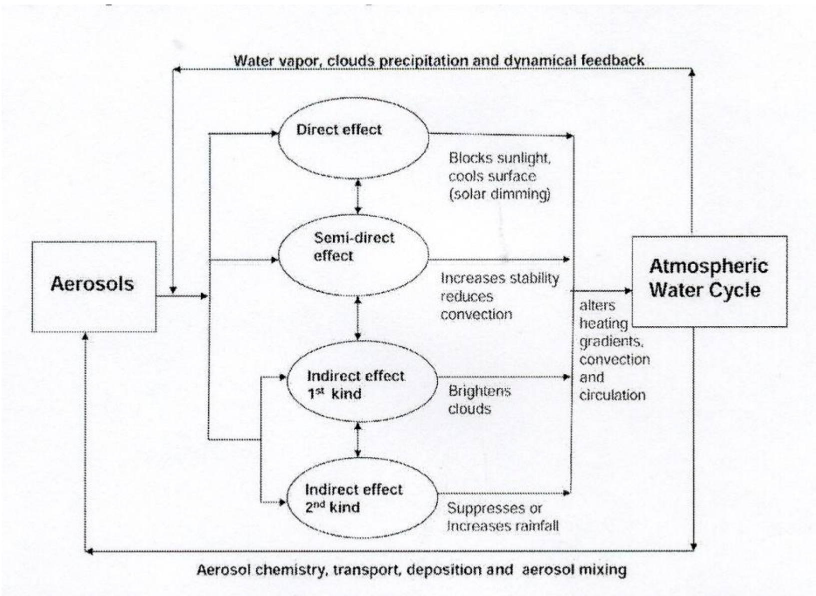


Figure 3.7: A schematic representation of the interaction of atmospheric water and aerosols.[47]

Something of the complexity of the situation may be gathered from the fact that if aerosols reduce ambient surface temperatures there should be an equivalent reduction in the intensity of the summer monsoon for two reasons. First, as we have seen, the summer monsoon is driven by the temperature differential between the warming land surface of the TP and the relatively cooler Indian Ocean. Therefore a reduction in temperature contrast between sea and land will reduce wind velocity, reducing wind friction at sea surface, an important agent of evaporation. Second, within a given temperature range, the amount of moisture the atmosphere can hold is reckoned to decrease by ~6% for each 1°C drop in temperature.

However, it is suggested that the summer monsoon westerlies have increased in velocity over recent years.[48] For, although rainfall in northern India has shown a decreasing trend for July, August and September in recent decades, June has shown an increasing trend. This may, in part,

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[43] (The area weighted mean radius of cloud droplet, or, more simply droplet size).
[44] Lau, K.M. *et al.* 2008.
[45] Jin, M. & Shepherd, J. 2008. Their introduction provides an excellent review of these anomalies.
[46] Shrestha A.B. *et al.* 2000 p.8315.
[47] Reproduced from Lau, K.M. *et al*. 2008 p.371.
[48] Goes *et al.* 2005.

result from reduced Himalayan snow cover and increased tropospheric warming in May.[49] The extent to which these anomalies are the result of aerosol forcing is the subject of considerable debate.

We now need to look at the positive temperature anomalies in the air column above the ABCs:

> The absorption of solar radiation by dust heats up the elevated surface air over slopes. On the southern slopes, the atmospheric heating is reinforced by black carbon from local emission. The heated airs rises via dry convection, creating a positive temperature anomaly in the mid-to-upper troposphere over the TP relative to the region in the south. In May through early June in a manner akin to an "elevated heat pump", the rising hot air forced by the increasing heating of the upper troposphere draws in warm and moist air over the Indian subcontinent, setting the stage for the onset of the South Asian summer monsoon.[50]

Thus Lau and colleagues proposed that the natural temperature differential that drives the monsoon is enhanced by aerosol forcing to effect an 'elevated heat pump' (EHP) to the extent that its positive forcing was greater than the negative forcing at the surface. In a further paper, published later in the same year, they proposed that this effect would bring the northern sector of the monsoon forward into May-June because of the enhanced circulation caused by peaking AOD in May. This anomalous increase in precipitation is thought to be balanced by a reduction for the rest of the monsoon. Others consider that pre monsoon precipitation is being enhanced even earlier in the year (March-April-May) at the expense of the months following. [51] Attractive as the EHP proposition is, it has been criticised by scientists working at the University of Maryland as having a very limited basis in observation. They conclude instead:

> The possibility that both aerosol and precipitation anomalies, in turn, are shaped by a slowly evolving, large scale circulation pattern cannot presently be ruled out, in part because current models and observational analyses are unable to tease apart regional feedbacks form the large-scale influence. [52]

Nepal is sandwiched between the industrial giants India and China. India is thought to produce about 10% of the world's black carbon,[53] 40% of which is derived form biomass burning.[54] 'Recent studies show that the sulphate aerosol burden in the atmosphere in Nepal is largely of Indian origin.' Yet there is no indication of an overall trend in precipitation in Nepal. Nepal's rainfall generally accords with that of northern (rather than all-) India's. [55] This may in part be the result of the monsoon reaching the eastern IGP before it reaches the western IGP (see Figure 3.3), where aerosol loading is heavier.[56]

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[49] Gautam, R. *et al.* 2009.
[50] Lau, K. *et al*. 2006, p.855.
[51] See Meehl *et al.* 2008.
[52] Nigam & Bollasina, 2010, p.7.
[53] As at 1996. Uncertainties apply to this estimate. See Bond *et al*. 2004, p.28.
[54] Ramanathan, V. & Carmichael, G. 2008.
[55] Shrestha A.B. *et al.* 2000. The quotation is from p. 321. Se also Xu, B. *et al.* 2009.
[56] Prasad, A. *et al*. 2005. See also Gautam, R. *et al.* 2009, p.3694.

A key factor in the three dimensional aerosol loading in Nepal is the country's complex topography. It is hardly surprising therefore that satellite CALIPSO-borne lidar measurements have shown aerosols at 5km asl:

> Either via direct and, or indirect radiative forcing and, or via cloud microphysics, the strong gradient in aerosol concentration appears therefore to be linked to the variability of orographic precipitation along the southern slopes of the Himalayas, thus collocating two of the world's steepest gradients of optical depth and topography in a region critical for the Asian summer monsoon.
>
> …all field studies agree on the consistent presence of high concentrations of aerosols at high altitude in locations far away from urban sources. [57]

We will return to the forcing effects of aerosols in Nepal under 3.6.2, when we consider Himalayan temperature increases. But for the moment it appears that the effects of aerosol forcing on the Indian monsoon may vary at a local and / or regional scale. We have already noted their differing functions as CCNs over ocean and over land. If this is indeed the case it is hardly surprising that considerable uncertainties surround aspects not only of their forcing capacity but their relationship to atmospheric circulation.

3.4.1.2 El Niño and the Indian Ocean Dipole

An important variable affecting the monsoon is (the) El Niño Southern Oscillation (ENSO). Normally the upwelling of deep cold water against the western side of equatorial South America leaves the continental seaboard arid. The water is warmed by being carried westward across the Pacific; associated easterly winds gather moisture, depositing rain on Indonesia and Australia. The ~5 year reversal of this system (ENSO) has considerable consequences upon world weather systems. These include warming across south and south east Asia including the Bay of Bengal.

Modelling by Barnett and colleagues[58] suggest that a heavy Eurasian snowfall (with consequential weaker monsoon (see 3.3.2)) should be considered among a number of factors contributing to the initiation of a weaker ENSO.

There appears to be room for discussion as to the extent to which ENSO events influence the weakening of the monsoon or are reciprocating this anomaly.

> El Niño warm phases are associated with warmer sea surface temperature in the equatorial Pacific as well as in the Indian Ocean, resulting in a decreased land–ocean thermal contrast, thus reducing the strength of the [summer] monsoon.[59]

Under ENSO overall rainfall is reduced, affecting western Nepal the most, with reductions of streamflow by perhaps 10%. The impact is noticeable but less severe in the east and least severe in central Nepal, [60] although pre monsoon and monsoon rainfall was reduced by approximately 15% during 1992–3 compared with the preceding decadal average at one station in the Likhu

[57] Shrestha, P. & Barros, A. 2010, p.8306.
[58] Barnett, T. et al. 1989.
[59] Shrestha, A.B. *et al.* 2000 p. 322.
[60] Shrestha, A. & Kostaschuk, R. 2004.

Khola basin (north west of Kathmandu) during a prolonged El Niño event.[61] However this should also be seen in the context of the Mount Pinatubo eruption the previous year, which reduced the world's atmospheric temperatures during the years in question. El Niño periods correlate with late-peaking low-magnitude runoff regimes.[62] It has been suggested that this is

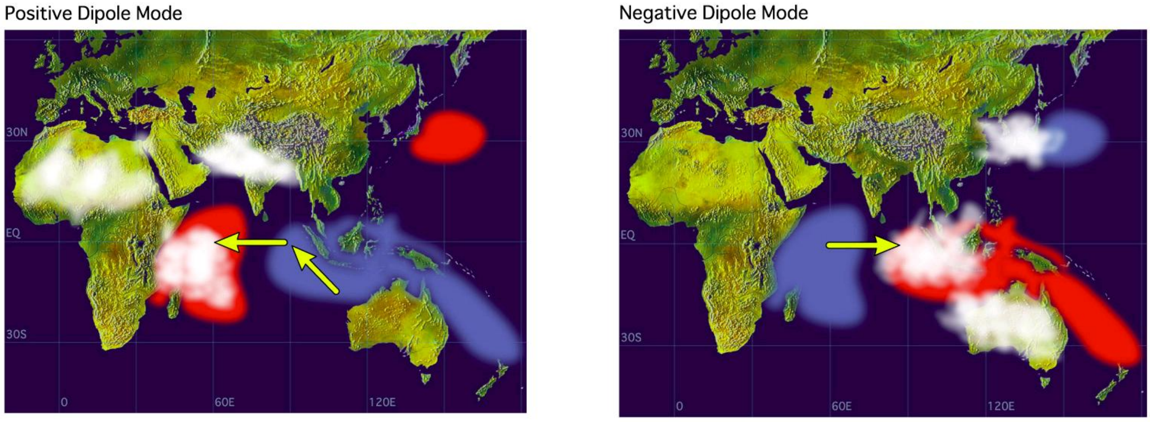

Figure 3.8: SST anomalies (red=warm, blue=cold). White areas represent increased convective activity with associated rainfall, yellow arrows wind direction.[63]

now a weakening trend and that since late 1970s Indian summer monsoon connection with ENSO has been attenuated thanks to the strengthening poleward jetstream over the North Atlantic.[64] Perhaps this is why the prolonged ENSO event of 1991–5 did not cause the expected drought in India.

The Indian Ocean dipole (IOD) is a periodic anomaly created by SST differences between the western and eastern equatorial Indian Ocean. A positive IOD is created between a warmer SSTs in the western equatorial Indian Ocean and cooler SSTs in the south eastern equatorial Indian Ocean and is associated with an enhanced monsoon rainfall.

Ahok's[65] studies of the relationship between ENSO and IOD concluded that where positive IOD and ENSO events occurred simultaneously, ENSO's negative effect on the summer monsoon was reduced.

3.5 Factors affecting winter precipitation in the central Himalayas

We have seen (3.2) that winter precipitation is higher in the Karakoram and Western Himalayas than in summer. This precipitation is carried by WDs as an extra-tropical frontal system that loses some of its characteristics while moving eastward across Afghanistan and Pakistan.[66] However occasionally this storm track can extend further east, depositing considerable quantities of precipitation as the airflow is forced upwards across the southern face of the Himalayas.

61 Gardner, R. & Gerrard, A. 2003.
62 Wagener, T & Franks, S. 2005.
63 JAMSTEC
64 Fowler, H. & Archer, D. 2005.
65 Ashok *et al.* 2004.
66 Hatwar, H. *et al.* 2005.

During one of these events in the winter of 1985 – 6 winter precipitation at Kyangchen (Langtang) at 3920m was nearly double that at Kathmandu airport (1336m), while in summer it as slightly less than half.[67] Although inter-annual variability of winter precipitation is large, this kind of event has occurred ten times between 1921 and 1986. However taken on a decadal average, winter precipitation will never exceed that of the summer in this region.

3.6 Changes to temperature regimes: Introduction

Since the 1960s temperatures in the Himalayas and the TP are found to have risen as fast as anywhere on the globe. Yet no such dramatic increases in temperature are found in the air column in the non-mountainous troposphere. We therefore need to examine possible reasons why the warming effect is so dramatically enhanced in high mountain regions.

Discussion of a potential enhanced relationship between altitude and temperature (altitude-dependent warming, ADW) has been discussed by Swiss and American scientist since the mid-1990s at least.[68] From the research that has followed, it is now evident that there is a strong positive correlation between increases in altitude and temperature (up to certain altitudes) in the TP and Himalayas, as in the Swiss Alps. However, it is important to keep in mind that, despite similarities of result, the causes of Alpine and Himalayan temperature increases are unlikely to be identical.

In the central and eastern Himalayas an observable cause for this warming lies in changes in precipitation and consequential loss of albedo. Increases in temperature can alter snowfall patterns, increasing the elevation at which precipitation falls as rain, lifting the snow line and limiting the duration of snowpack into the spring. Earlier snowmelt will reduce albedo allowing the surface to absorb radiation. Cloud albedo feedback is considered another important factor. GHGs are important drivers in increasing water vapour. Liu's[69] modelling shows how GHG forcing increases cloudiness (and downward long wave radiation) at lower altitudes (1500–2000m). Radiation reflected from the cloud's upper surface may create a temperature inversion, warming both atmosphere and hill slope above the cloud layer, which in turn will greatly affect the depth of the snow pack. To this extent ADW may be seen as the product of positive feedback.

While GHGs are clearly of enormous importance in driving temperature changes in the Himalayas, we also have to return to the subject of aerosols to consider their potential importance as drivers of environmental change. In this context we consider three distinct effects: *(i)* aerosol-enhanced reflective warming of the middle to upper troposphere, *(ii)* reducing the snow water equivalent (SWE) in snow crystals and *(iii)* the reduction of albedo from snow and ice surfaces.

(i) During March, dramatically increasing ABC density over the Indian Ocean at altitudes of 1km creates a doubling of AOD (from 0.18 to 0.4) which enhances atmospheric warming to a peak annual mean of 0.4°C at altitudes of between 2 and 4km. Over the Himalaya-Hindu-Kush

[67] Seko & Takahashi 2005.
[68] See, for example, Giorgi, F. *et al.* 1997.
[69] Liu, X. *et al*. 2009 See 3.6.1(*ii*).

warming will be carried vertically into the mid troposphere from March to September enhancing the peak annual mean by 0.6°C. Ramanathan and colleagues consider that ABCs enhance lower atmospheric heating by about 50%. 'The combined simulated warming of greenhouse gasses and ABCs at higher levels (3–5km asl) is about 1.2K or about 0.25K per decade during the period from 1950 to the present …'[70]

Prasad and colleagues[71] demonstrated that in the Himalayas and TP, which occupy the middle troposphere (4 – 7km), the annual warming trend has been 1.44°C over 30 years (1977 – 2008). Their findings show that warming is heavily loaded on the months of December to May, winter through the first part of the summer[72]. This period is associated with a longer residence period of anthropogenic aerosols in the atmosphere and, as we have seen, receives a boost from natural high mineral content from dust storms during April to June. The warming trend peaks during May, a process which goes into reverse in June. Such aerosol-forced winter warming is has profound consequences not only for snow fall but also for snow pack duration.

(ii) If aerosols reduce cloud effective radius, this is equally true for both liquid and solid precipitation. Aerosols also 'reduce the snow particle rime growth, resulting in lower snow water equivalent… Properly represented aerosols in climate models will apparently also work together with increasing temperature to reduce snow/ice in regions where heavy air pollution exists.'[73]

Monsoon rainfall can of course 'scavenge' or wash aerosols from the troposphere. But as we have seen in the Shrestha & Barros citation (3.4.1.1.) the Himalayas provide a barrier to the airflow 'collocating two of the world's steepest gradients of optical depth and topography.'[74] Ming and colleagues[75] discovered concentrations of black carbon in ice cores on the East Rongbuk Glacier in Everest's north east saddle. They were able to demonstrate pretty clear evidence that black carbon deposits were carried to this altitude from the summer, rather than winter, monsoon.[76] They discovered quite varied aerosol forcing over 50-year period (1950–2000) period, but on a five year smoothed average this was never less than 1 W m^{-2} increasing in the early 1970s and late 1990s to double that rate. Prasad and colleagues, examining ice cores form the Dasouopu Glacier (central Himalayas) found that deposited (anthropogenically derived) sulphates had doubled since the 1970s, another indication that aerosols are carried deep into the Himalayas.

(iii) Loss of albedo from aerosol pollution is another major consideration. It is well-known from the work of Østrem (1959) that a surface layer of fines on the face of a glacier or snowfield will affect its albedo. Student experiments are often undertaken to verify the effects of different

[70] Ramanathan, V. *et al*. 2007 pp. 576–7. It is important to note that experiments were conducted over the ocean off south west India.
[71] Prasad, A. *et al.* 2009.
[72] It is worth noting that this is at odds with Tmax findings of Shrestha *et al.*, (1999), who suggest that the burden of annual warming in the Himalayas is loaded onto the post-monsoon period. (See 3.6.2).
[73] Barnett T. et al. 2005 p. 307.
[74] See the NASA website for excellent simulation of this effect from CALIPSO-derived data.
[75] Ming, J. *et al.* 2008.
[76] Xu, B. *et al.* 2009 also found that aerosol deposits on some TP glaciers derived from the south. However it is important to note that there are also other directional influences across the TP.

debris cover depths on ablation. Yet many of these experiments use rock or fines with a high proportion of reflective silicate particles.

Ming and colleagues[77] discuss the density distribution of aerosols in the TP and surrounding areas, concluding that black carbon could reduce albedo by as much as 5%. Such would prove highly absorbent to solar radiation, and would have the potential to enhance ablation rates, even if the particles were several orders of magnitude smaller than those employed in the Østrem experiment. The presence of aerosols within ice cores can only derive from accumulations in the snowpack from either China or the Indian subcontinent. Their impact on the snowpack is well described by Xu and colleagues who show that black carbon concentrations significantly alter the albedo of a thick snow layer. [78]

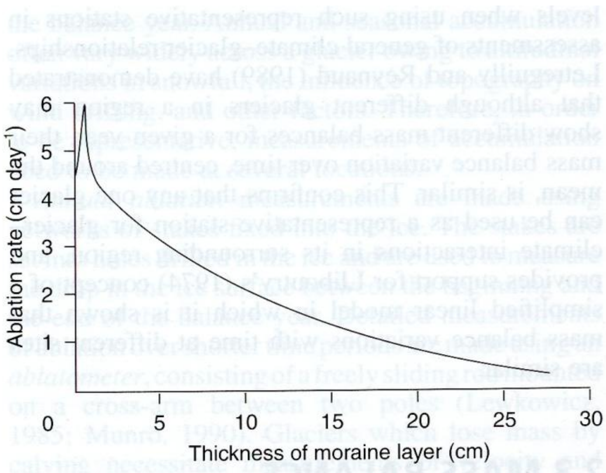

Figure 3.9: The effect of debris cover on surface ablation on a glacier.[79]

> The visible albedo of deep fresh snow, about 0.9 – 0.97, is decreased by 0.01 – 0.04 by a black carbon amount of 10 ng g^{-1} thus increasing absorption (1 minus albedo) of visible radiation by 10 – 100% depending on the size and shape of the snow crystals and on whether the soot is incorporated within the snow crystals or externally mixed. The impact of albedo change is magnified in the spring at the start of the melt season, because it allows the melt to begin earlier. Then as melting snow tends to retain some aerosols the surface concentration of black soot increases, and BC [black carbon] becomes even more effective at increasing melt of snow and ice.

Temperatures both in the TP and the Himalayas are increasing at apparently unprecedented rates. There may be overlaps in the cause of this, but since they are also subject to different weather systems they should be considered separately.

3.6.1 The Tibetan Plateau

The pioneering work that established the temperature/altitude correlation on the Tibetan Plateau was that of the Chinese scientists, Liu & Chen.[80] They built on existing research which suggested that the Tibetan Plateau is highly sensitive to changes in climate both from the palaeo record and from more recent research, particularly where snow/albedo feedback is concerned. They established that warming was largely commensurate with altitude reaching positive decadal trends

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[77] Ming, J. *et al.* 2009.
[78] Xu, B. *et al.* 2009. The following quotation is taken from p.22115, references omitted.
[79] Benn & Evans 1998 p. 73 modified from Østrem 1959.
[80] Liu, X. & Chen, B. 2000.

of ~0.25°C – 0.35°C at altitudes between ~3 – 4000m. The trend is clear if not even. Winter temperatures have risen faster than summer temperatures.

Liu and colleagues published further research in 2009 to include data from higher weather stations. It showed greater increases of temperature with altitude than the previous paper, particularly in winter where a decadal positive temperature trend of 0.8 °C was evident at altitudes of 4500 – 5000m with an annualised trend of 0.61°C at this altitude.[81]

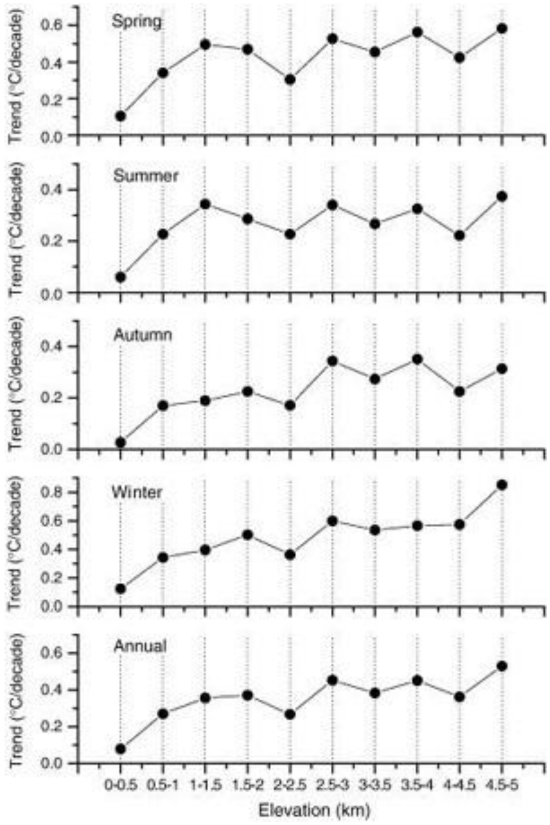

Figure 3.10 Comparison of linear trends of warming with altitude[82] Linear trends of mean minimum temperature as plotted for 10 elevation zones (1961 – 2006) on the basis of surface meteorological observations in the Eastern Tibetan Plateau.

There appear to be at least three factors contributing to altitude-dependent warming in the TP.

(i) Positive feedback caused by the loss of albedo from reduction in snowpack.

(ii) Cloud radiation feedback: The day time dispersal of low level cloud enhances the incoming shortwave radiation, which, with reduction of albedo, allows insolation to increase. GHG driven alterations in both the altitude and extent of cloud cover will vary levels of long wave 'outgoing' radiation. Night-time low level cloud now covers up to 60% of the TP, enhancing atmospheric counter radiation. (*i.e.* 'trapping' randomly directional long wave radiation within the envelope

[81] Liu, X. *et al.* 2009.

[82] Liu, X. *et al.* 2009. pp. 167. In a personal communication (25 April 2012) Zhi-Young Yin, co-author of this paper, points out that the pronounced 'kink' at 2 – 2.5km shows a slightly reduced warming rate to about 0.3°C/decade and suggests that various physical processes, including cloud formation could influence the trend in this zone. However because there were only 4 stations at this elevation, the findings are statistically weakened. The authors prefer to attach greater significance to the modelled results in this paper where the 'kink' is reduced.

between cloud and earth.) This also has the effect of reducing DTR by increasing temperatures at the lower end of the range by a greater amount than the warming at the upper end.[83]

Liu also points to earlier numerical studies that show that increases in CO_2 have the propensity to increase cloud cover at lower levels and decrease cloud cover at higher altitudes. 'The inflection point to switch from increasing trends to decreasing trends is found at 1.5 – 2km. In fact the decreasing (cloud cover) trend becomes more prominent as elevation increases, until it stabilises at around 4000m.'[84] His model further demonstrates the projected correlation between increases in absorbed solar radiation and reduction in depth of the snowpack up to about 4500m.

(iii) The altitudinal variation of aerosol types and densities (see 3.4.1). Clearly we are at the early stages of research into what appears to be a driving force in TP environmental issues.

3.6.2 Nepal

While some of the meteorological records available to Chinese scientists in the TP had a reasonably long historical reach, Nepal is less well served. Nepal has (or had in 1996) 264 functioning weather stations. Their distribution is uneven and only a third of them had been functioning for 30 years, the minimum term suggested by the World Meteorological Organisation as being appropriate for reliable meteorological mapping purposes, at the time when Nepal's first modern climatic and hydrological atlas was published (with Japanese support) in 1996.[85] Nevertheless, the publication provided an accessible data base through which Nepal's extraordinary climatic variations can be approached. It is tragic that funds are not available for this atlas to be updated on a decadal basis.

A second stepping stone in understanding Nepal's climate was provided by Arun Shrestha and colleagues in 1999 in a paper discussing temperature trends in Nepal 1971 – 94. The records show the mid 1970s as the point in which a cooling trend (identified from weather stations set up in the early 1960s) changed to a warming trend in all five regions of the country. This paper is probably the one most cited when it comes to discussing climate change in the Himalayas. The authors remain agnostic as to whether '[c]limatic changes in the Himalayan region could be a reflection of large scale climate changes, or [whether] they could even be driving them'.[86] Their hypothesis is that the summer monsoon exercises the most important control on temperatures in Nepal but they decided to focus on the mean of maximum temperatures (Tmax) because to establish changes in the mean diurnal temperature range (a more robust measure) would involve mean winter minimum temperatures (Tmin). There were concerns that these could be distorted by urbanisation and because '[a] preliminary analysis showed high year-to-year fluctuation and lack of significant trend in minimum temperature in most of the station records in Nepal.[87] This brings sharply into focus the difficulty of gathering and assessing the quality of data in remote locations. Their findings are summarised in the following table.

[83] Duan, A. & Wu, G. 2006.
[84] Liu, X. *et al.* 2009 p. 170.
[85] Chalise, S. *et al*. 1996.
[86] Shrestha, A.B. et al. 1999 p.2775.
[87] Shrestha, A.B. et al. 1999 p.2776.

Table 3.1: Regional mean temperature (max) trends for Nepal 1977 – 94. Units in degrees Celsius *per annum*.

	Seasonal				
Regions	Winter (Dec – Feb)	Pre monsoon (March – May)	Monsoon (June – Sept)	Post monsoon (Oct –Nov)	Annual (Jan – Dec)
Trans Himalaya	0.124[a]	0.005	0.109[b]	0.099[c]	0.09[b]
High Mountains and Himalaya	0.09[b]	0.05	0.062[b]	0.075[a]	0.057[b]
Mid Mountains	0.059[c]	0.05	0.055[b]	0.094[b]	0.075[b]
Siwalik	0.015	0.01	0.021	0.077[a]	0.041[a]
Terai	0.006	-0.004	0.014	0.069[a]	0.041[a]
All Nepal	0.061[a]	0.032	0.051[a]	0.081[b]	0.059[b]

[a] $p=\geq 0.01$; [b] $p \geq 0.001$; [c] $p \geq 0.05$ [88]
Source: Shrestha *et al.* 1999 p. 2781.

We have seen from Liu and colleagues (cited above) that loss of albedo from the reduced duration of snowpack can increase temperatures. Normally one would expect this process to be strongest in the post-winter, pre-monsoon period. But an important finding of this paper is that that the highest mean average seasonal temperature increases fall post-monsoon, leading one to wonder whether, in the case of the High Himalaya, this might result from decreases in monsoon precipitation as snowfall.

Apart from the trans-Himalayas, as far as altitudinal temperature increases are concerned, the highest rates of warming are found in the Middle Mountains, followed by the high mountains. The Siwaliks are also experiencing warming, but at a lower rate. The authors see this as a potential function of cloudiness and the advection of moist air, which is more saturated in the south than the north of the county.

It is possible to speculate that the enhanced warming in the post-monsoon period might also result from increasing anthropogenic aerosols. Natural aerosol densities would normally be much lower than their June-July-August peak at this time of year but industrial aerosols may be increasing proportionally.

Basing conclusions on the analysis of Tmax data alone may (in some part at least) account for the differences in seasonal loading aspect of annual warming trends shown in this paper and that of Prasad *et al.* 2009 above (3.6). There may be other explanations however, as Prasad noted broadly similar trends on the eastern side of the IGP.

Further decadal data analysis of Nepal's meteorological and hydrological temperatures is currently being undertaken. The rapid winter warming of the Middle Mountains may have little

[88] '*p*' values are intended to give an indication of the likelihood of a value occurring by accident, with *p*=1 meaning 100% likely and *p*=0.01 meaning one chance in a hundred.

influence upon seasonal rainfall patterns and high rainfall variability has been apparent for 30 years. By way of example, one mid hills temperature station No 1206 near Salleri[89] has recorded that average winter Tmins have remained well above 5°C (refrigerator temperature) since 2001 (in the 6 – 7°C range). In the previous two decades Tmins consistently dipped to 5°C or below, despite some prolonged warmer years in the early part of the series. DTR has actually increased because of consistently higher Tmax since 1992. At present winter precipitation (from nearby station 1204) shows no clear trend with a wide range of variability between ~0 and 40mm with occasional records of over 80mm over the same time scale. While the results from one weather station can only be described as very weakly indicative, they do demonstrate an accelerating process of winter warming.[90]

In 2007 Japanese scientist Fukui and colleagues[91] published a paper suggesting that mean annual temperatures were increasing faster in the Himalayas than on the TP. In making this difficult assessment they discovered the rate of permafrost retreat on a south-facing slope in Khumbu was in the order of 100 – 300m in 30 years. Obtaining data at altitudes of 5000 – 5500m is a considerable achievement, although consideration might have been given to the relative differences between the south-facing Himal and the flatter lying plateau. Nevertheless the findings are not at odds with those of Shresha and colleagues (1999), but further experiments are needed to corroborate the findings at an altitude where data gathering is (and probably always will be) problematic.

The role played by albedo is complex in a mountain environment. Deforestation has also been suggested as a driving force in raising temperatures in the Himalayas. Normally dense forest canopy has higher albedo than the exposed terraced earth, however this is not inevitable. We will see in the next chapter that the term 'forest' is applied to a range of tree densities depending upon who is using the term. Furthermore, some exposed earth surfaces have high albedo, depending on their parent geology. Deforestation is not now happening across the Himalayas at a rate where it is likely to affect temperatures. Indeed much of the Himalayan deforestation took place before the industrial revolution and this is discussed in the next chapter. For the moment, as Shrestha[92] points out, the greatest changes in land use (in the Siwaliks and Terai) are associated with the least changes in temperature.

[89] See Chalise, S. *et al.* 1996.
[90] This information from the DHM was used in an unpublished baseline survey report for The Glacier Trust.
[91] Fukui, K. *et al*. 2007.
[92] Shrestha, A.B. *et al*. 1999. For a discussion of this issue see p.2784.

Figure 3.12: A spring afternoon in the Siwaliks, March 2009. (Photo: The Glacier Trust)

From the discussions above concerning aerosols (3.4.1.1 & 3.6), it would be difficult to consider climatic change in Nepal's mountains without the implication that they have a major role in temperature forcing. It may well be that increased atmospheric pollution at lower altitudes (Siwaliks and Teri) together with the effect of 'bright clouds' may play a part in slowing the rate of temperature increase relative to the rest of the Himalayas. 'Slash and burn' or shifting agriculture is still practised in the Siwaliks, despite being discouraged. 'Pockets' of pollution, resulting from topographic features may be causing some of the temperature anomalies referred to in the previous chapter. Nevertheless, considerably more work needs to be done before clear causal links can be established.

Figure 3.11: Solu Khumbu. Left: ABC clearly visible above the Inkhu Khola valley looking south. Right: reverse angle looking north to Mera Peak. Photos by Susan Kaspari taken at about 4300m.[93]

[93] NASA (b).

3.7 Chapter summary

At the beginning of chapter 1 we noted a range of climatic effects listed by the IPCC as being characteristic of the effects of GHG-induced global warming over the past century. Amongst these effects, increased temperatures, reduced DTR, longer drought periods and increased heavy precipitation events will be familiar to many but by no means all Himalayan farmers. However, it would be a mistake simply to attribute all these aspects of climate change to GHGs. The Karakorum summers are now colder than they were and it has been noted, for example, that summer monsoon variability was greater ninety years ago than it has been since. ENSO's power to deliver periodic droughts may be weakening.

It is evident that aerosols have the capacity to affect weather patterns, but quantifying their capacity to alter or delay the summer monsoon as it affects Nepal will remain problematic, particularly after the monsoon season has started reducing atmospheric aerosol concentrations. While a strong correlation between atmospheric pollution and ADW is evident, at the time of writing little, if any, research on cloud radiation feedback in the Himalayas seems to have been published. Further work on the relationship between ADW and aerosol forcing at a local to regional scale is urgently needed.

This chapter therefore raises more questions than answers. It may be some years before sufficient data are available to give a clearer picture. The consequences from a development perspective are that, since we only have a partial view of the forces driving climatic change, greater emphasis must be placed on the potential variability of future climate scenarios. While a relatively common trend is increased periods of drought with shorter more intense precipitation events (Chapter 2), this is by no means universal. For lack of trend, particularly in relation to non-summer monsoon rainfall is also a relatively common 'trend'. Development interventions will therefore need to envisage a wider range of situations. This will place an increasing stress on aid resources, particularly where donors require rapid returns on their munificence.

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