7 THE HUMAN IMPACT ON CLIMATE AND THE ATMOSPHERE

World climates

The climate of the world is now known to have fluctuated frequently and extensively in the three or so million years during which humans have inhabited Earth. The bulk of these changes have nothing to do with human intervention. Climate has changed, and is currently changing, because of a wide range of different natural factors which operate over a variety of timescales (Figure 7.1).

No completely acceptable explanation of climatic change has ever been presented, and no one process acting alone can explain all scales of climate change. The complexity of possibly causative factors involved is daunting.

The complexity becomes evident if we follow the pathway of radiation derived from the ultimate force of climate – the sun. First of all, for reasons such as the varying tidal pull being exerted on the sun by the planets, the quality and quantity of solar radiation may change. It has been recognized that radiation from the sun changes both in quantity (through association with familiar phenomena such as **sunspots**, which are dark regions of lower surface temperature on the surface of the sun) and in quality (through changes in the ultraviolet range of the solar spectrum). Cycles of solar activity have been established for the short term by many workers, with 11- to 22-year cycles being noted in particular. Sunspot cycles of 80–90 years have also been postulated. The observations of sunspots in historical times have also given a measure of solar activity and one very striking feature is the near absence of sunspots between AD 1640 and 1710, a period sometimes called the Maunder Minimum. It is perhaps significant that this minimum occurred during some of the more extreme years of the inclement Little Ice Age.

The receipt of such varying radiation at Earth's surface might itself vary because of the presence of fine interstellar matter (nebulae) through which the earth might from time to time pass. This would tend to reduce the receipt of solar radiation. Likewise, the passage of the solar system through a dust lane bordering a spiral arm of the Milky Way galaxy might cause a temporary reduction in receipt of radiation output from the sun.

The receipt of incoming radiation will also be affected by the position and configuration of Earth. Such changes do take place, and there are three main

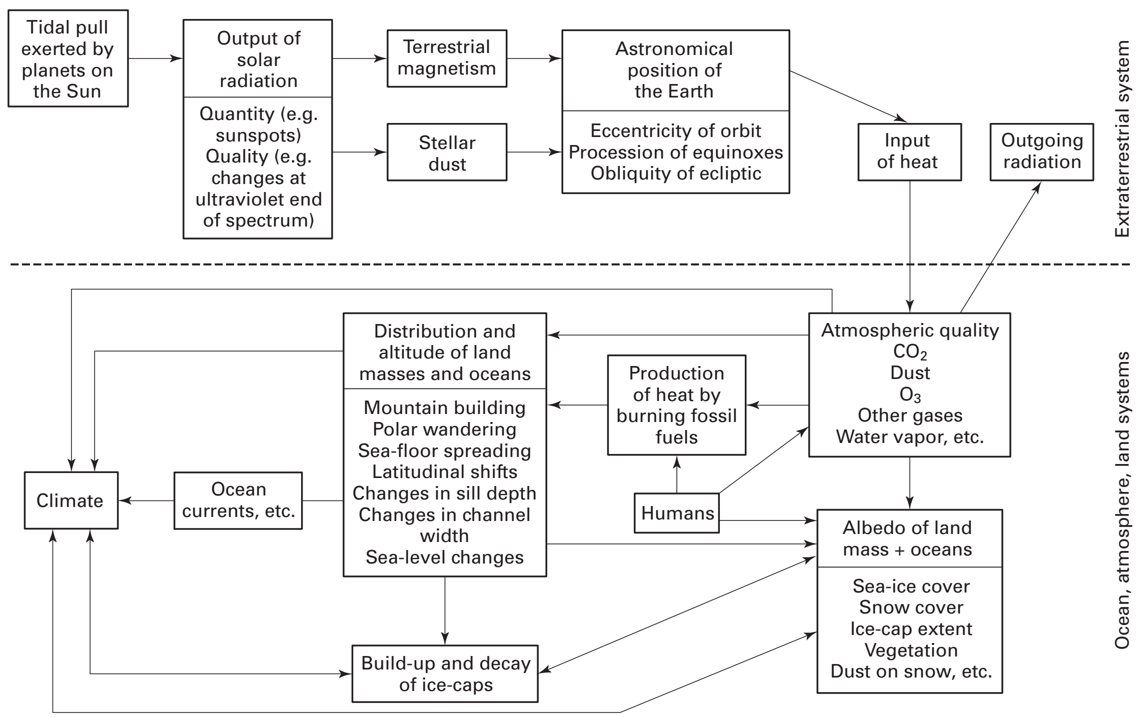

Figure 7.1 A schematic representation of some of the possible influences causing climatic change (after Goudie, 1992, figure 1).

astronomical factors which have been identified as of probable importance, with all three occurring in a cyclic manner. Firstly, Earth's orbit around the sun is not a perfect circle but an ellipse. If the orbit were a perfect circle then the summer and winter parts of the year would be equal in length. With greater eccentricity the length of the seasons will display a greater difference. Over a period of about 96,000 years, the eccentricity of Earth's orbit can 'stretch' by departing much further from a circle and then revert to almost true circularity.

Secondly, changes take place in the 'precession of the equinoxes', which means that the time of year at which the earth is nearest the sun varies. The reason is that Earth wobbles like a child's top and swivels round its axis. This cycle has a periodicity of about 21,000 years.

Thirdly, changes occur, with a periodicity of about 40,000 years, in the 'obliquity of the elliptic' – the angle between the plane of Earth's orbit and the plane of its rotational equator. This movement has been likened to the roll of a ship with a tilt varying from 21°39′ to 24°36′. The greater the tilt, the more pronounced is the difference between winter and summer.

These three cycles comprise what is often called the Milankovitch or Orbital Theory of climatic change. They have a temptingly close similarity in their periodicity to the durations of climatic change associated with the many glacials and interglacials of the past 1.6 million years. Indeed, they have been termed the 'pace-maker of the ice ages'.

Once the incoming solar radiation reaches the atmosphere, its passage to the surface of Earth is controlled by the gases, moisture, and particulate matter that are present. Essential importance has been attached to the role of dust clouds emitted from volcanoes. These could increase the backscattering of incoming radiation and thus promote cooling. Volcanic dust veils produced by, for example, the eruption of Krakatoa in the 1880s and by Mount Pinatubo in 1991 caused global cooling for a matter of a few years. However, changing levels of volcanic activity are not the only way in which changes in atmospheric transparency might occur. For example, dust can be emplaced into the atmosphere by the wind erosion of fine-grained sediment and soil, and we know from the extensive deposits of wind-laid silts (loess) of glacial age that during the glacial maxima the atmosphere was probably very dusty, contributing to global cooling.

Carbon dioxide, methane, nitrous oxide, sulfur dioxide, and water vapor can also modify the receipt of solar radiation. Particular attention has focused in recent years on the role of carbon dioxide (CO_2) in the atmosphere. This gas is virtually transparent to incoming solar radiation but absorbs outgoing terrestrial infrared radiation – radiation that would otherwise escape to space and result in heat loss from the lower atmosphere. In general, through the mechanism of this so-called greenhouse effect, low levels of CO_2 in the atmosphere would be expected to lead to cooling, and high levels would be expected to produce a 'heat trap'. The same applies to levels of methane and nitrous oxide, which, molecule for molecule, are even more effective greenhouse gasses than CO_2. Recently it has proved possible to retrieve CO_2 from gas bubbles preserved in layers of ice in deep ice-cores drilled from the polar regions. Analyses of changes in CO_2 concentrations in these cores have provided truly remarkable results and have demonstrated that CO_2 changes and climatic changes have progressed in approximate synchroneity over the past 800,000 years (EPICA, 2004). Thus the last interglacial around 120,000 years ago was a time of high CO_2 levels, and the last glacial maximum around 18,000 years ago of low CO_2 levels, and the early Holocene a time of very rapid rise in CO_2 levels. The reasons for the observed natural change in greenhouse gas concentrations are still the subject of active scientific research.

Once incoming radiation from the sun reaches the surface of Earth it may be absorbed or reflected according to the nature of the surface, and in particular according to whether it is land or water, covered in vegetation or desert, and whether it is mantled by snow.

The effect of the received radiation on climate also depends on the distribution and altitude of landmasses and oceans. These too are subject to change in a wide variety of ways – the plates that comprise Earth's crust are ever moving, mountain belts may grow or subside, and oceans and straits open and close. These processes shift areas into new latitudes, transform the world's wind belts, and modify the climatically very important ocean currents.

In this discussion of causes it is also crucial to consider feedbacks. Such feedbacks are responses to the original forcing factors that act either to increase or intensify the original forcing (we call this positive feedback) or to decrease or reverse it (negative feedback). Clouds, ice and snow, and water vapor are three of the most important feedback mechanisms. An example of a positive feedback is the role of snow. Under cold conditions this falls rather than rain, it changes the **albedo** (reflectivity) of the ground surface and causes further cooling of the air above it.

Figure 7.2 Air pollution in Cape Town, South Africa. The combustion of fossil fuels, including coal, to generate electricity and to power vehicles, is a major cause not only of local air pollution but also of the increase in greenhouse gas loadings in the atmosphere.

Finally, it may well be that the atmosphere and the oceans possess a degree of internal instability that furnishes a built-in mechanism of change so that some small and random change might, through the operation of positive feedbacks and the passage of thresholds, have extensive and long-term effects. Small triggers might have large consequences.

While humans are at present incapable of modifying some of these natural mechanisms of climate change – the output of solar radiation, the presence of fine interstellar matter, Earth's orbital variations, volcanic eruptions, mountain building, and the overall pattern of land masses and oceans – there are some key areas where humans may be capable of making significant changes to global climates (Figure 7.2). The most important categories of influence are in terms of the chemical composition of the atmosphere and the albedo of Earth's surface, although, as the following list of *possible mechanisms* shows, there are some others that also need to be considered.

1 Gas emissions
- CO_2 – industrial and agricultural
- methane
- chlorofluorocarbons (CFCs)
- nitrous oxide
- krypton 85
- water vapor
- miscellaneous trace gases

2 Aerosol generation
- dust, smoke, etc.

3 Thermal pollution

4 Albedo change
- dust addition to ice caps
- deforestation
- overgrazing

5 Extension of irrigation

6 Alteration of ocean currents by constricting straits

7 Diversion of freshwaters into oceans, affecting thermohaline circulation system

8 Impoundment of large reservoirs

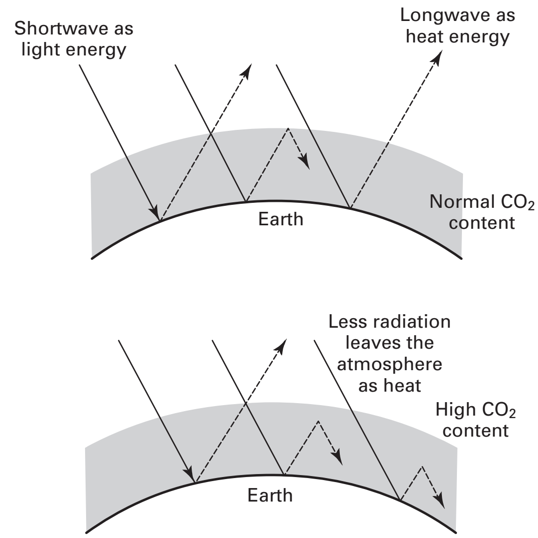

Figure 7.3 The greenhouse effect: shortwave radiation from the sun is absorbed at Earth's surface, which in turn radiates heat at far longer wavelengths because of its temperature of around 280 K, compared with 6000 K for the sun.

The greenhouse gases

Carbon dioxide

The greenhouse effect occurs in the atmosphere because of the presence of certain gases that absorb infrared radiation (Figure 7.3). Light and ultraviolet radiation from the sun are able to penetrate the atmosphere and warm Earth's surface. This energy is re-radiated as infrared radiation, which, because of its longer wavelength, is absorbed by certain substances such as water vapor, carbon dioxide, and other trace gases. This causes the average temperature of Earth's surface and atmosphere to increase. Should the quantities of each substance in the atmosphere be increased, then the greenhouse effect will become enhanced (Hansen et al., 1981).

In reality the term 'greenhouse effect' is something of a misnomer. As Henderson-Sellers and Robinson (1986: 60) explain:

> We know that a greenhouse maintains its higher internal temperature largely because the shelter it offers reduces the turbulent transfers of energy away from the surface rather than because of any radiative considerations. Thus while the greenhouse effect remains valid, and vital, for the atmosphere it might be better to think of the physical processes in terms of the 'leaky bucket' analogy. . . . Here an increase in the amount of gas with absorption bands in the infrared part of the spectrum is represented by a decrease in the size of the hole in the bucket. The surface temperature, represented by the depth of the water in the bucket, rises as more absorbing gases enter the atmosphere.

There are a number of ways in which humans have been enhancing the greenhouse effect, the most significant of which is to increase carbon dioxide levels in the atmosphere. This may have started with deforestation in the Holocene (Ruddiman, 2003) but has accelerated in recent centuries. Since the beginning of the industrial revolution humans have been taking stored carbon out of the earth in the form of coal, petroleum, and natural gas, and burning it to make carbon dioxide (CO_2), heat, water vapor, and smaller amounts of sulfur dioxide (SO_2) and other gases. The pre-industrial level of carbon dioxide is a matter of some debate, but may have been as low as 260–70 ppm by volume (Wigley, 1983). The present level is over 370 ppmv, and the upward trend is evident in records from various parts of the world (Figure 7.4). At the present rate it would reach 500 ppmv by the end of the twenty-first century. The prime cause of increased

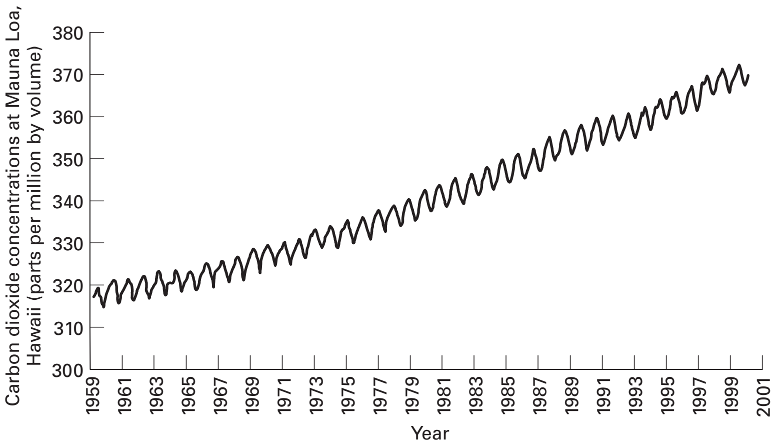

Figure 7.4 Carbon dioxide concentrations at Mauna Loa, Hawaii.

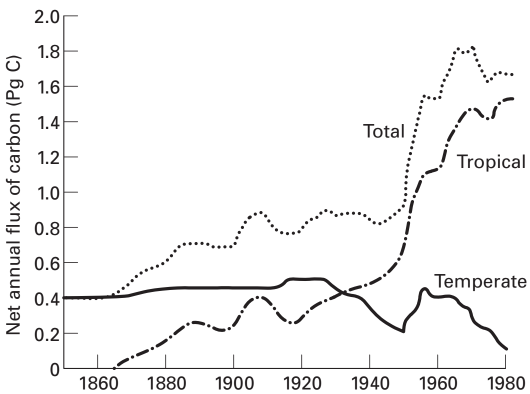

Figure 7.5 The net annual flux of carbon from deforestation in tropical and temperate zones globally, 1850–1980 (after Houghton and Skole, 1990, figure 23.2, and Woodwell, 1992, figure 5.2).

carbon dioxide emissions is fossil fuel combustion and cement production (*c.* 5.5 ± 0.5 Gt C year^{-1} in the 1980s), but with the release of carbon dioxide by changes in tropical land use (primarily deforestation) being a significant factor (*c.* 1.6 ± 1.0 Gt C year^{-1}). The amount of carbon derived from deforestation has increased greatly from about 0.4 Gt C year^{-1} in 1850 (Figure 7.5) (Woodwell, 1992). The various relationships between land-use change and the build up of greenhouse gases are reviewed by Adger and Brown (1994).

Other gases

In addition to carbon dioxide, it is probable that other gases will contribute to the greenhouse effect (Table 7.1). Individually their effects may be minor, but as a group

Table 7.1 Sources of four principal greenhouse gases*

Gas	*Natural sources*	*Human-derived sources*
Carbon dioxide	Terrestrial biosphere Oceans	Fossil fuel combustion Cement production Land-use modification
Methane	Natural wetlands Termites Oceans and freshwater lakes	Fossil fuels (natural gas production, coal mines, petroleum industry, coal combustion) Enteric fermentation (e.g., cattle) Rice paddies Biomass burning Landfills Animal waste Domestic sewage
Nitrous oxide	Oceans Tropical soils (wet forests, dry savannas) Temperate soils (forests, grassland)	Nitrogenous fertilizers Industrial sources Land-use modification (biomass burning, forest clearing) Cattle and feed lots
Chlorofluorocarbons†	Nil	Rigid and flexible foam Aerosol propellants Teflon polymers Industrial solvents

*Sources listed in order of decreasing magnitude of emission except where otherwise indicated.
†Sources of chlorofluorocarbons not in order of decreasing magnitudes of emission.

Table 7.2 Radiative forcing relative to CO_2 per unit molecule change in the atmosphere. Source: extracted from Houghton et al. (1990: 53, table 2.3)

Gas	*Relative radiative forcing*	*Residence time in atmosphere (years)*
CO_2	1	100
CH_4 (methane)	21	10
N_2O (nitrous oxide)	206	100–200
CFC-11	12,400	65
CFC-12	15,800	130

they may be major (Ramanathan, 1988). Indeed, molecule for molecule some of them may be much more effective as greenhouse gases than CO_2, as the data in Table 7.2 show.

One of the more important of the trace gases is methane (CH_4), which has a strong infrared absorption band at 7.66 μm. Ice-core studies and recent direct observations (Figure 7.6b) suggest that until the beginning of the industrial revolution in the eighteenth century background levels were relatively stable at around 600 parts per billion by volume (ppbv) although they may have been increased prior to that by rice farming and other agricultural activities (Ruddiman and Thomsen, 2001). They rose steadily between AD 1700 and 1900, and then increased still more rapidly, attaining levels that averaged 1300 ppbv in the early 1950s and 1600 ppbv by the mid-1980s (Khalil and Rasmussen, 1987) and over 1700 ppbv in the 1990s. This increase of 2.5 times over background levels results primarily from increased rice cultivation in waterlogged paddy fields, the enteric fermentation produced in the growing numbers of flatulent domestic cattle, and the burning of oil and natural gas (Crutzen et al., 1986).

Chlorofluorocarbons (CFCs), despite their relatively trace amounts in the atmosphere, have increased very markedly in terms of their emissions (Figure 7.6c) and their concentrations in recent decades, resulting from their use as refrigerants, foam makers, fire control agents, and propellants in aerosol cans. They have a very strong greenhouse effect even in relatively small amounts. On the other hand, the ozone depletion they have caused in the stratosphere may to some limited extent counteract this effect, for stratospheric ozone depletion results in a decrease in **radiative forcing** (Houghton et al., 1992). Conversely, the build up of lower-level **tropospheric** ozone can contribute to the greenhouse effect.

Nitrous oxide (N_2O) is also no laughing matter, for it can contribute to the greenhouse effect, primarily by absorption of infrared at the 7.8 and 17 μm bands. Combustion of hydrocarbon fuels, the use of ammonia-based fertilizers, deforestation, and biomass burning are among the processes that could lead to an increase in atmospheric N_2O levels (Figure 7.6a and Table 7.3b). Atmospheric N_2O concentrations have increased from around 275 ppbv in pre-industrial times to 311 ppbv in 1992.

Other trace gases that could play a greenhouse role include bromide compounds, carbon tetrafluoride, carbon tetrachloride, and methyl chloride.

The continued role of greenhouse gases other than CO_2 in changing the climate is already not greatly less important than that of CO_2. If present trends continue, the combined concentrations of atmospheric CO_2 and other greenhouse gases would be radiatively equivalent to a doubling of CO_2 from pre-industrial levels possibly as early as the 2030s. The relative amounts of radiative forcing for different greenhouse gases since pre-industrial times are, according to the Intergovernmental Panel on Climate Change (IPCC, 1996), as follows:

CO_2	1.56 W m^{-2}
CH_4	0.47 W m^{-2}
N_2O	0.14 W m^{-2}
CFCs and HCFCs	0.25 W m^{-2}
tropospheric ozone	0.40 W m^{-2}

Another important feature of the various greenhouse gases is their residence time in the atmosphere. Methane has a residence time of about 10 years, the shortest of all the greenhouse gases. This means that if we could stop the enhanced emissions of that gas, its concentration in the atmosphere should fall to its natural level in a decade. By contrast N_2O (100–200 years) and CO_2 (*c.* 100 years) have much longer residence times, so that even if we could control their sources immediately it would still take a very long time for them to fall to their natural levels.

Global temperatures have been climbing since the end of the nineteenth century (Figure 7.7) and it is now regarded as highly probable that increased greenhouse gas loadings in the atmosphere have contributed to this.

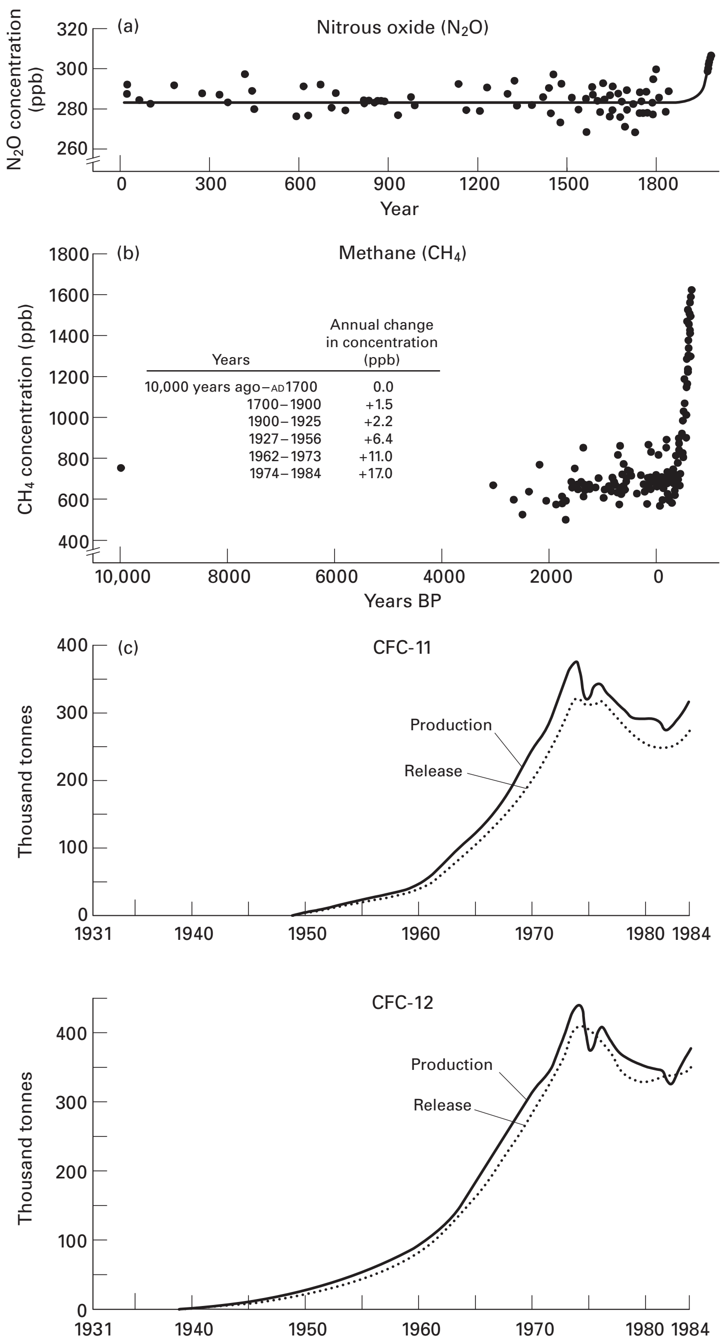

Figure 7.6 The changing concentrations of accessory greenhouse gases in the atmosphere: (a) nitrous oxide – note these remained fairly constant between 23,000 years ago and AD 1850 at approximately 285 parts per billion (after Khalil and Rasmussen, 1987); (b) methane (after Khalil and Rasmussen, 1987); (c) the changing production and release of two CFC gases (CFC-11 and CFC-12) between 1931 and 1992.

Aerosols

Aerosols are finely divided solid or liquid particles dispersed in the atmosphere. In general terms it is believed that they affect climate because they intercept and scatter a small portion of incoming radiation from the sun, thus reducing the energy reaching the ground. This is called the 'direct effect' of aerosols. The direct effect increases with both the number and size of aerosols in the atmosphere. Aerosols also have an 'indirect effect'. This is because they are key elements in cloud formation. The number of cloud droplets that a cloud possesses is determined by the number of aerosols available on to which water vapor can

Table 7.3 (a) Estimates of source strengths and sinks. Source: data in UNEP (1991)
(a) Methane (CH_4)

Sources/sinks	*Best estimate (10^6 t $year^{-1}$)*	*Range (10^6 t $year^{-1}$)*
Sources:		
(a) natural wetlands	115	100–200
(b) rice paddies	110	25–170
(c) enteric fermentation (animals)	80	65–100
(d) gas drilling, venting, transmission	45	25–50
(e) biomass burning	40	20–80
(f) termites	40	10–100
(g) landfills	40	20–70
(h) coal mining	35	19–50
(i) oceans	10	5–20
(j) freshwaters	5	1–25
(i) CH_4 hydrate destabilization	5	0–100
Sinks:		
(a) removal by soils	30	15–45
(b) reaction with OH	500	400–600
(c) atmospheric increase	44	40–48

(b) Nitrous oxide (N_2O)

Sources/sinks	*Range (10^6 t $year^{-1}$)*
Sources:	
(a) oceans	1.4–2.6
(b) soils (tropical forests)	2.2–3.7
(c) soils (temperate forests)	0.7–1.5
(d) fossil fuel combustion	0.1–0.3
(d) biomass burning	0.02–0.2
(e) fertilizer (including groundwater)	0.01–2.2
Sinks	
(a) removal by soils	Unknown
(b) photolysis in the stratosphere	7–13
(c) atmospheric increase	3–4.5

condense. Clouds with large numbers of droplets reflect more sunlight back into space and so can also contribute to cooling. Aerosols do not, however, inevitably cause cooling (Weave et al., 1974). So, for example, Idso and Brazel (1978) and Brazel and Idso (1979) point to the two contrasting tendencies of dust: the backscattering effect producing cooling, and the thermal-blanketing effect causing warming. The second of these absorbs some of Earth's thermal radiation that would otherwise escape to space, and then re-radiates

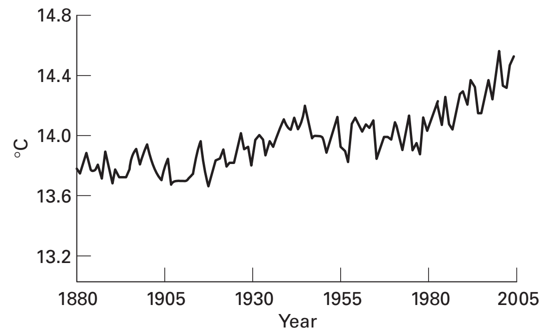

Figure 7.7 Global average temperature at Earth's surface, 1880–2002.

a portion of this back to the land surface, raising surface temperatures. They believe that natural dust from volcanic emissions tends to enter the stratosphere (where backscattering and cooling are the prime consequences), while anthropogenic dust more frequently occurs in the lower levels of the atmosphere, causing thermal blanketing and warming.

There are a variety of ways in which human activities have increased atmospheric aerosol loadings. One of these is industrial emission of smoke and dust particles (Davitya, 1969), though whether this is a matter of only local rather than regional or global significance is a matter for debate.

Industrialization is not, however, the sole source of particles in the atmosphere, nor is a change in temperature the only possible consequence. Bryson and Barreis (1967), for example, argue that intensive agricultural exploitation of desert margins, such as in Rajasthan, India, would create a dust pall in the atmosphere by exposing larger areas of surface materials to deflation in dust storms. This dust pall, they believe, would so change atmospheric temperature that convection, and thus rainfall, would be reduced. Observations on dust levels over the Atlantic during the drought years of the late 1960s and early 1970s in the Sahel suggest that the degraded surfaces of that time led to a great (threefold) increase in atmospheric dust (Prospero and Nees, 1977). There is thus the possibility that human-induced desertification generates dust which could in turn increase the degree of desertification by its effect on rainfall levels. Biomass burning of tropical savannas may also add large numbers of aerosols to the atmosphere. Studies of the links between dust in the atmosphere and temperatures,

Figure 7.8 The Gulf War of 1991 led to the deliberate release and burning of oil in Kuwait. Fears were expressed at the time that smoke palls might have regional and global climate effects. In general, subsequent research has suggested that such fears may have been exaggerated.

precipitation, and clouds is a major area of current research (see, e.g., Tegen et al., 1996; Miller and Tegen, 1998).

The most catastrophic effects of anthropogenic aerosols in the atmosphere could be those resulting from a nuclear exchange between the great powers. Explosion, fire, and wind might generate a great pall of smoke and dust in the atmosphere, which would make the world dark and cold. It has been estimated that if the exchange reached a level of several thousand megatons, a 'nuclear winter' would be as low as –15° to –25°C (Turco et al., 1983), although more recent simulations by Schneider and Thompson (1988) suggest that some previous estimates may have been exaggerated. They suggest that in the Northern Hemisphere maximum average land surface summertime temperature depressions might be of the order of 5–15°C. The concept is discussed by Cotton and Piehlke (1995, chapter 10).

Fears were also expressed that as a result of the severe smoke palls (Figure 7.8) generated by the Gulf War in 1991 there might be severe climate impacts. Studies have suggested that because most of the smoke generated by the oil-well fires stayed in the lower troposphere and only had a short residence time in the air, the effects were local (some cooling) rather than global, and that the operation of the monsoon was not affected to any significant degree (Browning et al., 1991; Bakan et al., 1991). Furthermore, in the event the emissions of smoke particles were less than some forecasters had predicted, and they were also rather less black (Hobbs and Radke, 1992).

Aircraft, both civil and military, discharge some water vapor into the atmosphere as contrails. At present, the water content of the stratosphere is very low, as is the exchange of air between the lower stratosphere and other regions. Consequently, comparatively modest amounts of water vapor discharge by aircraft could have a significant effect on the natural balance. It is possible that contrails and the development of thin cirrus clouds could lead to warming of the Earth's surface (IPCC, 1999).

Over the world's oceans a major source of aerosols is dimethylsulfide (DMS). This is produced by planktonic algae in seawater and then oxidized in the atmosphere to form sulfate aerosols. Because the albedo of clouds (and thus Earth's radiation budget) is sensitive to cloud-condensation nuclei density, any factor that controls planktonic algae may have an important impact on climate. The production of such plankton could be affected by water pollution in coastal areas or by global warming (Charlson et al., 1987). However, an even more important source of sulfate aerosols is the burning of fossil fuels and the subsequent emission of sulfur dioxide (SO_2) (Charlson et al., 1992), and these types of sulfate aerosol are concentrated over and downwind of major industrial regions. They have probably served to reduce the rate of global warming that has taken place in this century and may help to explain the cessation in global warming that took place in some regions between the 1940s and 1970s. Indeed, climate models that have predicted the amount of increase in global average temperature as a result of the rising concentrations of greenhouse gases have given a greater amount of temperature rise since the last century than has actually occurred. The newer climate models, which include the effect of these aerosols, produce predicted changes that have considerable similarity to the observed patterns of change (Taylor and Penner, 1994).

Vegetation and albedo change

Incoming radiation of all wavelengths is partly absorbed and partly reflected. Albedo is the term used to describe the proportion of energy reflected and

Table 7.4 Albedo values for different land-use types. Source: from miscellaneous data in Pereira (1973) collated by author

Surface type	*Location*	*Albedo (%)*
Tall rain forest	Kenya	9
Lake	Israel	11.3
Peat and moss	England and Wales	12
Pine forest	Israel	12.3
Heather moorland	England and Wales	15
Evergreen scrub (maquis)	Israel	15.9
Bamboo forest	Kenya	16
Conifer plantation	England and Wales	16
Citrus orchard	Israel	16.8
Towns	England and Wales	17
Open oak forest	Israel	17.6
Deciduous woodland	England and Wales	18
Tea bushes	Kenya	20
Rough grass hillside	Israel	20.3
Agricultural grassland	England and Wales	24
Desert	Israel	37.3

hence is a measure of the ability of the surface to reflect radiation.

Land-use changes create differences in albedo which have important effects on the energy balance, and hence on the water balance, of an area. Tall rain forest may have an albedo as low as 9%, while the albedo of a desert may be as high as 37% (Table 7.4).

There has been growing interest recently in the possible consequences of deforestation on climate through the effect of albedo change. Ground deprived of vegetation cover as a result of deforestation and overgrazing (as in parts of the Sahel) has a very much higher albedo than ground covered in plants. This could affect temperature levels. Satellite imagery of the Sinai–Negev region of the Middle East shows an enormous difference in image between the relatively dark Negev and the very bright Sinai-Gaza strip area. This line coincides with the 1948–9 armistice line between Israel and Egypt and results from different land-use and population pressures. Otterman (1974) has suggested that this albedo change has produced temperature changes of the order of 5°C.

Charney and others (1975) have argued that the increase in surface albedo, resulting from a decrease in plant cover, would lead to a decrease in the net incoming radiation, and an increase in the radiative cooling of the air. Consequently, they argue, the air would sink to maintain thermal equilibrium by adiabatic compression, and cumulus convection and its associated rainfall would be suppressed. A positive feedback mechanism would appear at this stage, for the lower rainfall would in turn adversely affect plants and lead to a further decrease in plant cover. However, this view is disputed by Ripley (1976) who suggests that Charney and his colleagues, while considering the impact of vegetation changes on albedo, have completely ignored the effect of vegetation on evapotranspiration. He points out that vegetated surfaces are usually cooler than bare ground since much of the absorbed solar energy is used to evaporate water, and concludes from this that protection from overgrazing and deforestation might, in contrast to Charney's views, be expected to lower surface temperatures and thereby reduce, rather than increase, convection and precipitation.

Removal of humid tropical rain forests has also been seen as a possible mechanism of anthropogenic climatic change through its effect on albedo. Potter et al. (1975) have proposed the following model for such change:

Deforestation
↓
Increased surface albedo
↓
Reduced surface absorption of solar energy
↓
Surface cooling
↓
Reduced evaporation and sensible heat flux from the surface
↓
Reduced convective activity and rainfall
↓
Reduced release of latent heat, weakened Hadley circulation and cooling in the mid and upper troposphere
↓
Increased tropical lapse rates
↓
Increased precipitation in the latitude bands 5–25°N and 5–25°S, and a decrease in the Equator–pole temperature gradient
↓
Reduced meridional transport of heat and moisture out of equatorial regions
↓
Global cooling and decrease in precipitation between 45° and 85°N and 40° and 60°S

However, some studies (e.g., Potter et al., 1981) suggested that globally over the past few thousand years

Table 7.5 Some recent studies of climatic effects of vegetation removal

Source	*Location*	*Subject*
Fuller and Ottka (2002)	West Africa	Albedo and desertification
Fu (2003)	East Asia	Reduced atmospheric and soil moisture in East Asian monsoon region
Chase et al. (2000)	Global	Effects on main circulation features
Werth and Avissar (2002)	Amazonia	Reduction of local precipitation, evapotranspiration and cloudiness and also global effects
Berbet and Costa (2003)	Amazonia	Precipitation variability
Reale and Zirmeyer (2000)	Mediterranean Basin	Increased precipitation prior to deforestation in Roman times
Taylor et al. (2002)	Sahel	Rainfall decrease

the climatic effects of albedo changes wrought by humans have been small and probably undetectable. Similarly, Henderson-Sellers and Gornitz (1984) sought to model the possible future effects of albedo changes produced by humans and also predicted that there would be but little alteration brought about by current levels of tropical deforestation. On the other hand, Lean and Warrilow (1989) used a general circulation model (GCM) which showed greater changes than previous models and suggested that Amazon basin deforestation would, through the effects of changes in surface roughness and albedo, lead to reductions in both precipitation and evaporation. Likewise a UK Meteorological Office GCM indicated that the deforestation of both Amazonia and Zaire would by changing surface albedo cause a decrease in precipitation levels (Mylne and Rowntree, 1992). There are now an increasing number of modeling experiments that suggest vegetation removal can have important regional and even global effects (Table 7.5), although there are considerable divergences between different models (Nobre et al., 2004). Nonetheless, most show *decreases* in mean evapotranspiration of from 25.5 to 985.0 mm per year, *increases* in mean surface temperatures of 0.1–3.8°C and *reductions* in regional precipitation. It is even possible that the effects of Amazonian deforestation on precipitation could extend some distance away from there, to the Dakotas and the Midwest Triangle in the USA, as modeling by Werth and Avissar (2002) has shown.

Albedo effects may be especially sensitive in higher latitudes as well. As Betts (2000) has pointed out, in a snowy environment, forests are generally darker than open land, because trees generally remain exposed when cultivated land can become entirely snow-covered. Snow-free foliage is darker than snow. This means that forest has a smaller surface albedo and so may exert a warming influence. Thus land cover changes in the boreal forest zone can have substantial climatic implications.

Forests, irrigation, and climate

The replacement of forest with crops, in addition to leading to a change in surface albedo of the type just discussed, also changes some other factors that may have climatic significance, including surface aerodynamic roughness, leaf and stem areas, and amount of evapotranspiration (Betts, 2003). In particular large amounts of moisture may be transpired by deep rooting plants, which means that moisture is pumped back into the atmosphere, leading to increased levels of precipitation. Conversely, removal of such deep-rooting vegetation could exacerbate drought.

The belief that forests can increase precipitation levels has a long history (Thornthwaite, 1956; Grove, 1997), and it has been the basis of action programs in many lands. For example, the American Timber Culture Act of 1873 was passed in the belief that if settlers were induced to plant trees on the Great Plains and prairies, precipitation would be increased sufficiently to eliminate the climatic hazards to agriculture. On the other hand, at much the same time, the view was expressed that 'rain follows the plow'. Aughey, working in Nebraska, for example, believed that, after the soil is 'broken', rain as it falls is absorbed by the soil 'like a high sponge', and that the soil gives this absorbed moisture slowly back to the atmosphere by evaporation (cited by Thornthwaite, 1956: 569), and so increases the rainfall.

These two early and contradictory views illustrate the confusion that still surrounds this question today. Forests undoubtedly influence rates of evapotranspiration, the flow of streams, the level of groundwater, and microclimates, but there is little reliable evidence to suggest that regional rainfall is either significantly increased by forest or that attempts to augment rainfall levels on desert margins by widespread planting of forest belts are likely to achieve relatively much; the aridity of deserts and their margins is controlled dominantly by the gross features of the general circulation, especially the subsiding air associated with the large high-pressure cells of the subtropics.

Although forests may not necessarily have a proven effect on regional or continental rainfall levels, they are far more effective than other vegetation types at trapping other kinds of precipitation, especially cloud, fog, and mist. Hence deforestation or afforestation can affect water budgets through the degree to which they intercept nonrainfall precipitation.

There is one other land-use change that may result in measurable changes in precipitation; namely large-scale crop irrigation in semi-arid regions. The High Plains of the USA are normally covered with sparse grasses and have dry soils throughout the summer; evapotranspiration is then very low. In the past five decades irrigation has been developed throughout large parts of the area, greatly increasing summer evapotranspiration levels. Barnston and Schickdanz (1984) have produced strong statistical evidence of warm-season rainfall enhancement through irrigation in two parts of this area: one extending through Kansas, Nebraska, and Colorado, and a second in the Texas Panhandle. The largest absolute increase was in the latter area and, significantly, occurred in June, the wettest of the three heavily irrigated months. The effect appears to be especially important when stationary weather fronts occur, for this is a situation that allows for maximum interaction between the damp irrigated surface and the atmosphere. Hailstorms and tornadoes are also significantly more prevalent than over non-irrigated regions (Nicholson, 1988). However, Moore and Rojstaczer (2001) think that overall the irrigation effect is both difficult to quantify unambiguously and probably of minor significance.

Bonan (1997) has tried to model the climatic consequences of replacing the natural forests of the USA with crops and has argued that it would cause cooling of up to 2°C in the summer months over a wide region of the central USA. He suggests (p. 484) that 'land use practices that resulted in extensive deforestation in the Eastern USA, replacing forests with crop, have resulted in a significant climate change that is comparable to other well known anthropogenic climate forcings.'

The possible effects of water diversion schemes

The levels of the Aral and Caspian Seas in Central Asia have fallen, as have water tables all over the wide continental region. There have been proposals to divert some major rivers to help overcome these problems. However, this raises difficult questions 'because it appears to touch a peculiarly sensitive spot in the existing climatic regime of the northern hemisphere' (Lamb, 1977: 671). The low-salinity water which forms a 100–200 m upper layer to the Arctic Ocean is in part caused by the input of freshwater from the large Russian and Siberian rivers. This low-salinity water is the medium in which the pack ice at present covering the polar ocean is formed. The tapping of any large proportion of this river flow might augment the area of saltwater in the Arctic Ocean and thereby reduce the area of pack ice correspondingly. Temperatures over large areas might rise, which in turn might change the position and alignment of the main thermal gradients in the Northern Hemisphere and, with them, the jet stream and the development and steering of cyclonic activity. However, assessment of this particular climatic impact is still very largely speculative, and some numerical models indicate that the climate of the Arctic will not be drastically affected by river diversions (Semtner, 1984).

Lakes

It has often been implied that the presence of a large body of inland water must modify the climate around its shores, and therefore that artificial lakes have a significant effect on local or regional climates. Climatic changes produced by the construction of a reservoir are the result of a variety of factors (Vendrov, 1965): the creation of a body of water with a large heat capacity that reduces the continentality of the climate; the substitution of a water surface for a land surface and the rise of the groundwater level in the littoral zone supplying moisture to the evaporating surface (leading to a rise in wind velocity above the lake and in the littoral zone). Schemes have been put forward for augmenting desert rainfall by flooding desert basins in the Sahara, Kalahari, and Middle East (see, e.g., Schwarz, 1923). However, whether evaporation from lake surfaces can raise local precipitation levels is open to question, for precipitation depends more on atmospheric instability than upon the humidity content of the air. Moreover, most lakes are too small to affect the atmosphere materially in depth, so that their influence falls heavily under the sway of the regional circulation. In addition, one needs to remember that some of the world's driest deserts occur along coastlines. Thus a relatively small artificial lake would be even more impotent in creating rainfall (see Crowe, 1971: 443–50).

However, the climatic effect of artificial lakes is evident in other ways, notably in terms of a local reduction in frost hazard. In the case of the Rybinsk reservoir (*c*. 4500 km^2) in the CIS, it has been calculated that the climatic influence extends 10 km from the lake and that the frost-free season has been extended by 5–15 days on average (D'Yakanov and Reteyum, 1965).

Urban climates

One consequence of the burning of fossil fuels is the production of heat. This is probably most important on the local scale where it can be identified as the 'urban heat island'. On a broader scale, the amount of energy used by humans has been negligible compared both with the resources of solar energy and with the energy of photosynthesis of plants. In global terms the total amount of heat released by all human activity is roughly 0.01% of the solar energy absorbed at the surface (Kellogg, 1978: 215). Such a small fraction would have a negligible effect on the overall heat balance of Earth.

It has been said that 'the city is the quintessence of man's capacity to inaugurate and control changes in his habitat' (Detwyler and Marcus, 1972). One way in which such control becomes evident is in a study of urban climates (Landsberg, 1981). Individual urban areas can at times, with respect to their weather, 'have similar impacts as a volcano, a desert, and as an irregular forest' (Changnon, 1973: 146). Some of the changes that can result are listed in Table 7.6.

Table 7.6 Effects of cities on climate. Source: H. Landsberg in Griffiths (1976: 108)
(a) Average changes in climatic elements caused by cities

Element	*Parameter*	*Urban compared with rural (–, less; +, more)*
Radiation	On horizontal surface	–15%
	Ultraviolet	–30% (winter); –5% (summer)
Temperature	Annual mean	+0.7°C
	Winter maximum	+1.5°C
	Length of freeze-free season	+2 to 3 weeks (possible)
Wind speed	Annual mean	–20 to –30%
	Extreme gusts	–10 to –20%
	Frequency of calms	+5 to 20%
Relative humidity	Annual mean	–6%
	Seasonal mean	–2% (winter); –8% (summer)
Cloudiness	Cloud frequency + amount	+5 to 10%
	Fogs	+100% (winter); –30% (summer)
Precipitation	Amounts	+5% to 10%
	Days	+10%
	Snow days	–14%

(b) Effect of city surfaces

Phenomenon	*Consequence*
Heat production (the heat island)	Rainfall + Temperature +
Retention of reflected radiation by high walls and dark-colored roofs	Temperature +
Surface roughness increase	Wind – Eddying +
Dust increase (the dust dome)	Fog + Rainfall + (?)

Compared with rural surfaces, city surfaces (Table 7.6b) absorb significantly more solar radiation, because a higher proportion of the reflected radiation is retained by the high walls and dark-colored roofs of the city streets. The concreted city surfaces have both great thermal capacity and conductivity, so that

Table 7.7 Annual mean urban – rural temperature differences of cities. Source: from data in Detwyler (1971), and Wilby (2003)

City	*Temperature differences (°C)*
Chicago, USA	0.6
Washington, DC, USA	0.6
Los Angeles, USA	0.7
Paris, France	0.7
Moscow, Russia	0.7
Philadelphia, USA	0.8
Berlin, Germany	1.0
New York, USA	1.1
London, UK	1.8

heat is stored during the day and released by night. By contrast the plant cover of the countryside acts like an insulating blanket, so that rural areas tend to experience relatively lower temperatures by day and night, an effect enhanced by the evaporation and transpiration taking place. In addition, the energy partitioned for evapotranspiration is less in urban areas, leading to greater surface heating. Another thermal change in cities, contributing to the development of the 'urban heat island', is the large amount of artificial heat produced by industrial, commercial and domestic users.

In general the highest temperature anomalies are associated with the densely built-up area near the city center, and decrease markedly at the city perimeter. Observations in Hamilton, Ontario, and Montreal, Quebec, suggested temperature changes of 3.8 and 4.0°C km^{-1} respectively (Oke, 1978). Temperature differences also tend to be highest during the night. The form of the urban temperature effect has often been likened to an 'island' protruding distinctly out of the cool 'sea' of the surrounding landscape. The rural–urban boundary exhibits a steep temperature gradient or 'cliff' to the urban heat island. Much of the rest of the urban area appears as a 'plateau' of warm air with a steady but weaker horizontal gradient of increasing temperature towards the city center. The urban core may be a 'peak' where the urban maximum temperature is found. The difference between this value and the background rural temperature defines the *urban heat island intensity* (*Tu–r* (max)) (Oke, 1978: 225).

Table 7.7 lists the average annual urban – rural temperature differences for several large cities. Values range from 0.6 to 1.8°C. The relationship between city size and urban – rural difference, however, is not necessarily linear; sizeable nocturnal temperature contrasts have been measured even in relatively small cities. Factors such as building density are at least as important as city size, and high wind velocities will tend to eliminate the heat island effect.

Nonetheless, Oke (1978: 257) has found that there is some relation between heat-island intensity and city size. Using population as a surrogate of city size *Tu–r* (max) is found to be proportional to the log of the population. Other interesting results of this study include the tendency for quite small centers to have a heat island, the observation that the maximum thermal modification is about 12°C, and the recognition of a difference in slope between the North American and the European relationships (Figure 7.9). The explanation for this last result is not clear, but it may be related to the fact that population is a surrogate index of the central building density.

The relationship between maximum heat island intensity and urban population, the sky-view factor (a measure of building height and density), and the impermeable surface coverage of cities is shown in Figure 7.9. The heat island intensity increases with all three indices (Nakagawa, 1996).

In many older towns and cities in western Europe and North America, the process of 'counter-urbanization' has in recent years led to a decline in population, and it is worth considering whether this is reflected in a decline in the intensity of urban heat islands. One attempt to do this, in the context of London (Lee, 1992), revealed the perplexing finding that the heat-island intensity has decreased by day, but increased by night. The explanation that has been tentatively advanced to explain this is that there has been a decrease in the receipt of daytime solar radiation as a result of vehicular atmosphere pollution, whereas at night the presence of such pollution absorbs and re-emits significant amounts of outgoing terrestrial radiation, maintaining higher urban nocturnal minimum temperatures. The urban heat island may also be more marked at night because reduced nocturnal turbulent mixing keeps the warmer air near the surface. In mid-latitude cities such as London, urban heat island effects are generally stronger in summer than in winter because of higher levels of solar radiation being absorbed by building materials during the day. In winter the urban – rural

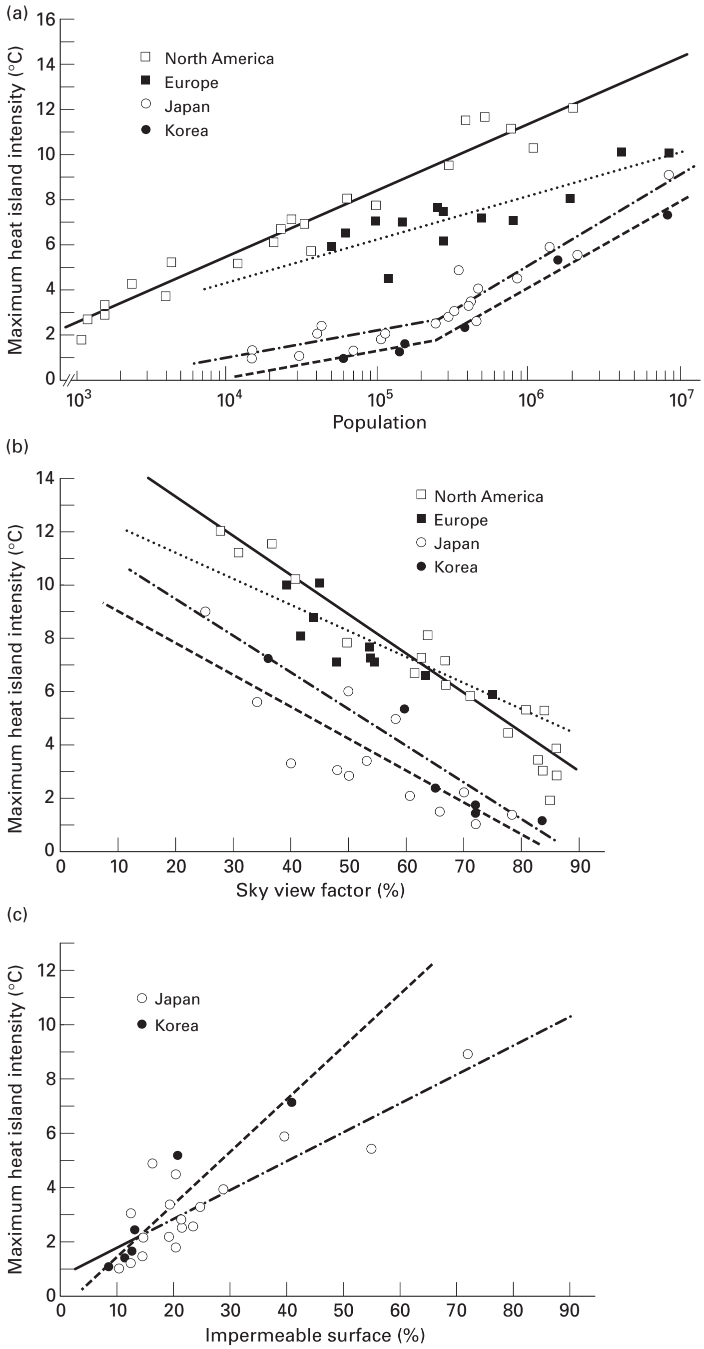

Figure 7.9 Relations between the maximum heat island intensity and (a) urban population for Japanese, Korean, North American, and European cities; (b) the sky view factor for Japanese, Korean, North American, and European cities; (c) the ratio of impermeable surface coverage for Japanese and Korean cities (after Nakagawa, 1996, figures 2, 3, and 4).

contrast is weaker because solar energy absorption is lower and hence there is less energy to radiate, despite higher levels of urban space heating (Wilby, 2003).

The existence of the urban heat island has a number of implications: city plants bud and bloom earlier, some birds are attracted to the thermally more favorable urban habitat, humans find added warmth stressful if the city is already situated in a warm area, during summer heatwaves exacerbated temperatures may cause mortality among sensitive members of the population, and less winter space-heating is required but, conversely, more summer air-conditioning is necessary.

The urban-industrial effects on clouds, rain, snowfall, and associated weather hazards such as hail and

Table 7.8 Areas of maximum increases (urban – rural difference) in summer rainfall and severe weather events for eight American cities. Source: after Changnon (1973: 144, figure 1.5)

City	*Rainfall*		*Thunderstorms*		*Hailstorms*	
	%	*Location**	*%*	*Location**	*%*	*Location**
St Louis	+15	B	+25	B	+276	C
Chicago	+17	C	+38	A, B, C	+246	C
Cleveland	+27	C	+42	A, B	+90	C
Indianapolis	0	–	0	–	0	–
Washington, DC	+9	C	+36	A	+67	B
Houston	+9	A	+10	A, B	+430	B
New Orleans	+10	A	+27	A	+350	A, B
Tulsa	0	–	0	–	0	–

*A = within city perimeter; B = 8–24 km downwind; C = 24–64 km downwind.

thunder are harder to measure and explain than the temperature changes (Darungo et al., 1978). The changes can be related to various influences (Changnon, 1973: 143):

- thermally induced upward movement of air;
- increased vertical motions from mechanically induced turbulence;
- increased cloud and raindrop nuclei;
- industrial increases in water vapor.

Table 7.8 illustrates the differences in summer rainfall, thunderstorms, and hailstorms between various rural and urban areas in the USA. These data indicate that in cities rainfall increases ranged from 9 to 27%, the incidence of thunderstorms increased by 10 to 42%, and hailstorms increased by 67 to 430%.

An interesting example of the effects of major conurbations on precipitation levels is provided by the London area. In this case it seems that the mechanical effect of the city was dominant in creating localized maxima of precipitation both by being a mechanical obstacle to air flow, on the one hand, and by causing frictional convergence of flow, on the other (Atkinson, 1975). A long-term analysis of thunderstorm records for southeast England is highly suggestive – indicating the higher frequencies of thunderstorms over the conurbation compared with elsewhere (Atkinson, 1968). The similarity in the morphology of the thunderstorm isopleth and the urban area is striking (Figure 7.10a and b). Moreover, Brimblecombe (1977) shows a steadily increasing thunderstorm frequency as the city has grown (Figure 7.10c).

Similarly the detailed Metromex investigation of St Louis in the USA (Changnon, 1978) shows that in the summer the city affects precipitation and other variables within a distance of 40 km. Increases were found in various thunderstorm characteristics (about +10 to +115%), hailstorm condition (+3 to +330%), various heavy rainfall characteristics (+35 to +100%), and strong gusts (+90 to +100%).

Two main factors are involved in the effect that cities have on winds: the rougher surface they present in comparison with rural areas; and the frequently higher temperatures of the city fabric.

Buildings, especially those in cities with a highly differentiated skyline, exert a powerful frictional drag on air moving over and around them (Chandler, 1976). This creates turbulence, with characteristically rapid spatial and temporal changes in both direction and speed. The average speed of the winds is lower in built-up areas than over rural areas, but Chandler found that in London, when winds are light, speeds are greater in the inner city than outside, whereas the reverse relationship exists when winds are strong. The overall annual reduction of wind speed in central London is about 6%, but for the higher velocity winds (more than 1.5 m per second) the reduction is more than doubled.

Studies in both Leicester and London (Chandler, 1976), England, have shown that on calm, clear nights, when the urban heat-island effect is at its maximum, there is a surface inflow of cool air towards the zones of highest temperatures. These so-called 'country breezes' have low velocities and become quickly decelerated by intense surface friction in the suburban areas. A practical implication of these breezes is that they transport pollution from the outer parts of an urban area into the city center, accentuating the pollution problem during smogs.

Urban air pollution

The concentration of large numbers of people, factories, power stations, and cars means that large amounts

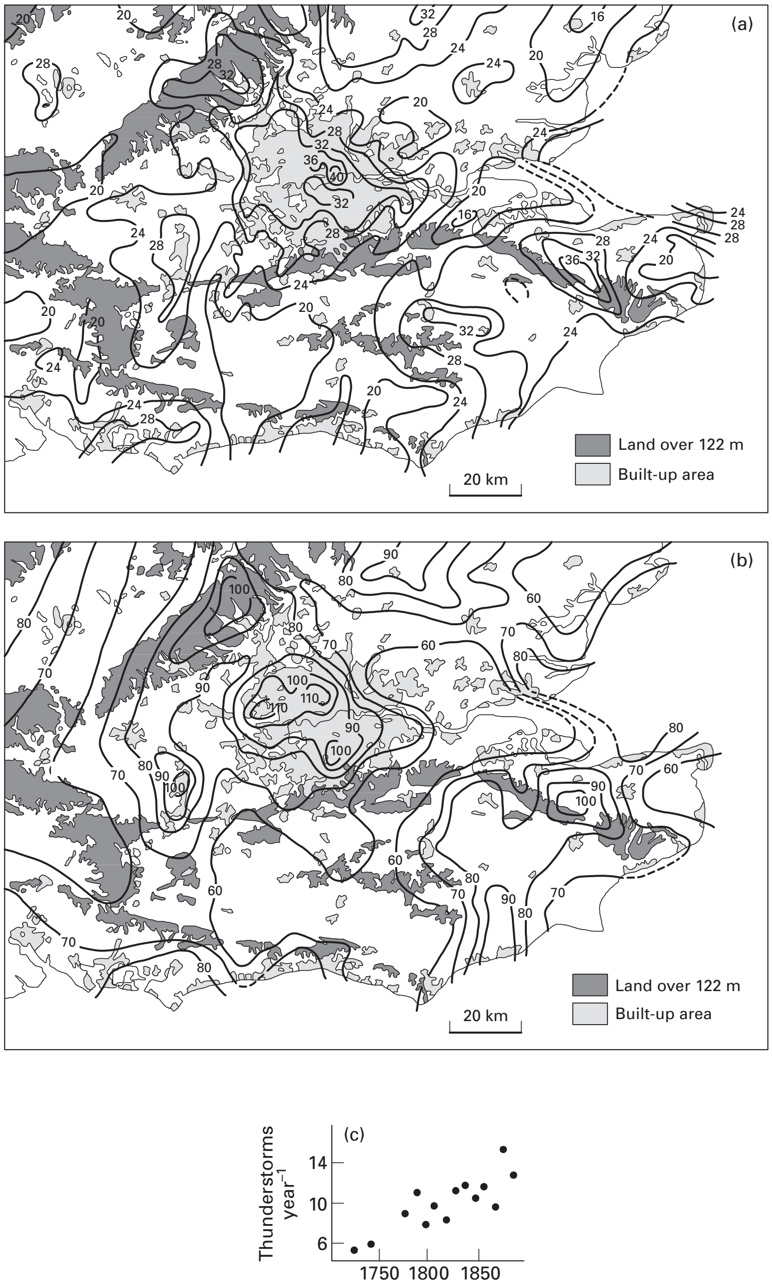

Figure 7.10 Thunder in southeast England: (a) total thunder rain in southeast England, 1951–60, expressed in inches (after Atkinson, 1968, figure 6); (b) number of days with thunder overhead in southeast England, 1951–60 (after Atkinson, 1968, figure 5); (c) thunderstorms per year in London (decadal means for whole year) (after Brimblecombe, 1977, figure 2).

of pollutants may be emitted into urban atmospheres. If weather conditions permit, the level of pollution may build up (Figure 7.11). The nature of the pollutants (Table 7.9) has changed as technologies have changed. For example, in the early phases of the industrial revolution in Britain the prime cause of air pollution in cities may have been the burning of coal, whereas now it may be vehicular emissions. Different cities may have very different levels of pollution, depending on factors such as the level of technology, size, wealth, and antipollution legislation. Differences may also arise because of local topographic and climatic conditions. Photochemical smogs, for example, are a more serious threat in areas subjected to intense sunlight.

Figure 7.11 In December 1952 the city of London was affected by severe smog. Visibility was reduced and smog-masks had to be worn out of doors. Many people with weak chests died. Since then, because of legislation, the incidence of smog has declined markedly.

The variations in pollution levels between different cities are brought out in Figure 7.12, which shows data for two types of pollution for a large range of city types. The data were prepared for the years 1980–4 by the Global Environmental Monitoring System of the United Nations Environmental Program (UNEP). Figure 7.12a shows concentrations of total particulate matter. Most of this comes from the burning of poor-quality fuels. The shaded horizontal bar indicates the range of concentrations the UNEP considers a reasonable target for preserving human health. Note that the annual mean levels range from a low of about 35 $\mu g\ m^{-3}$ to a high of about 800 $\mu g\ m^{-3}$: a range of about 25-fold! The higher values appear to be for rapidly growing cities in the developing countries. Some cities, however, such as Kuwait, may have unusually high values because of their susceptibility to dust storms from desert hinterlands. The lower values tend to come from cities in the developed world (e.g., western Europe, Japan, and North America).

Figure 7.12b shows concentrations for sulfur dioxide. Much of this gas probably comes from the burning of high-sulfur coal. Once again, the horizontal shaded bar indicates the concentration range considered by UNEP to be a reasonable target for preserving human health.

These data indicate that the concentrations of sulfur dioxide can differ by as much as three times among different sites within the same urban areas and by as much as 30 times between different urban areas.

Table 7.9 Major urban pollutants

Type	*Some consequences*
Suspended particulate matter (characteristically 0.1–25 μm in diameter)	Fog, respiratory problems, carcinogens, soiling of buildings
Sulfur dioxide (SO_2)	Respiratory problems, can cause asthma attacks. Damage to plants and lichens, corrosion of buildings and materials, production of haze and acid rain
Photochemical oxidants: ozone and peroxyacetyl nitrate (PAN)	Headaches, eye irritation, coughs, chest discomfort, damage to materials (e.g., rubber), damage to crops and natural vegetation, smog
Oxides of nitrogen (NO_x)	Photochemical reactions, accelerated weathering of buildings, respiratory problems, production of acid rain and haze
Carbon monoxide (CO)	Heart problems, headaches, fatigue, etc.
Toxic metals: lead	Poisoning, reduced educational attainments and increased behavioral difficulties in children
Toxic chemicals: dioxins, etc.	Poisoning, cancers, etc.

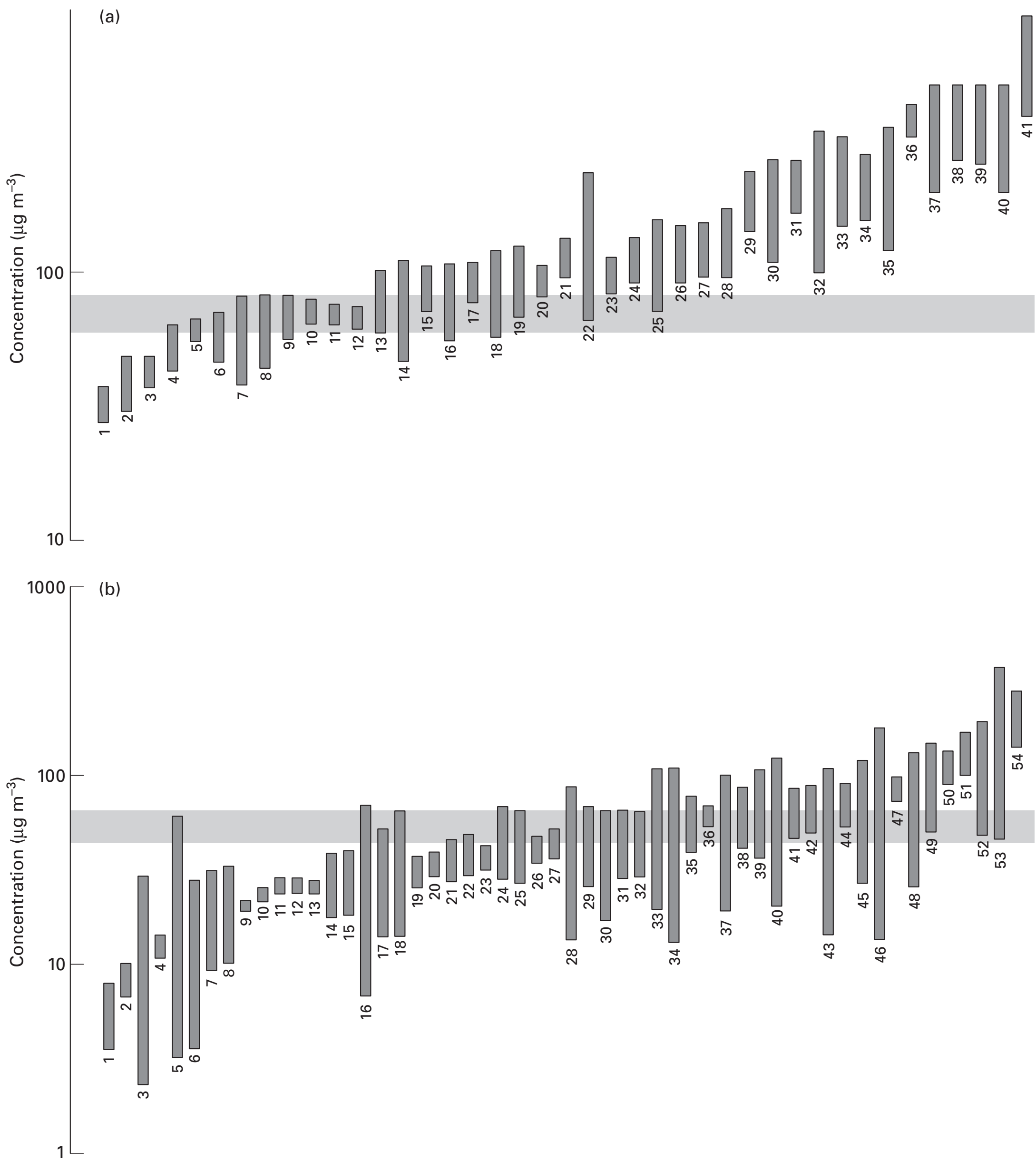

Figure 7.12 (a) The range of annual averages of total particulate matter concentrations measured at multiple sites within 41 cities, 1980–1984. The shading indicates the concentration range recommended by the United Nations Environment Program as a reasonable target for preserving human health. Each numbered bar represents a city, as follows: 1, Frankfurt; 2, Copenhagen; 3, Cali; 4, Osaka; 5, Tokyo; 6, New York; 7, Vancouver; 8, Montreal; 9, Fairfield; 10, Chattanooga; 11, Medellin; 12, Melbourne; 13, Toronto; 14, Craiova; 15, Houston; 16, Sydney; 17, Hamilton; 18, Helsinki; 19, Birmingham; 20, Caracas; 21, Chicago; 22, Manila; 23, Lisbon; 24, Accra; 25, Bucharest; 26, Rio de Janeiro; 27, Zagreb; 28, Kuala Lumpur; 29, Bombay; 30, Bangkok; 31, Illigan City; 32, Guangzhou; 33, Shanghai; 34, Jakarta; 35, Tehran; 36, Calcutta; 37, Beijing; 38, New Delhi; 39, Xi'an; 40, Shenyang; 41, Kuwait City. (b) The range of annual averages of sulfur dioxide concentrations measured at multiple sites within 54 cities, 1980–1984. Each numbered bar represents a city, as follows: 1, Craiova; 2, Melbourne; 3, Auckland; 4, Cali; 5, Tel Aviv; 6, Bucharest; 7, Vancouver; 8, Toronto; 9, Bangkok; 10, Chicago; 11, Houston; 12, Kuala Lumpur; 13, Munich; 14, Helsinki; 15, Lisbon; 16, Sydney; 17, Christchurch; 18, Bombay; 19, Copenhagen; 20, Amsterdam; 21, Hamilton; 22, Osaka; 23, Caracas; 24, Tokyo; 25, Wroclaw; 26, Athens; 27, Warsaw; 28, New Delhi; 29, Montreal; 30, Medellin; 31, St Louis; 32, Dublin; 33, Hong Kong; 34, Shanghai; 35, New York; 36, London; 37, Calcutta; 38, Brussels; 39, Santiago; 40, Zagreb; 41, Frankfurt; 42, Glasgow; 43, Guangzhou; 44, Manila; 45, Madrid; 46, Beijing; 47, Paris; 48, Xi'an; 49, São Paulo; 50, Rio de Janeiro; 51, Seoul; 52, Tehran; 53, Shenyang; 54, Milan. (Source: Graedel and Crutzen, 1993.)

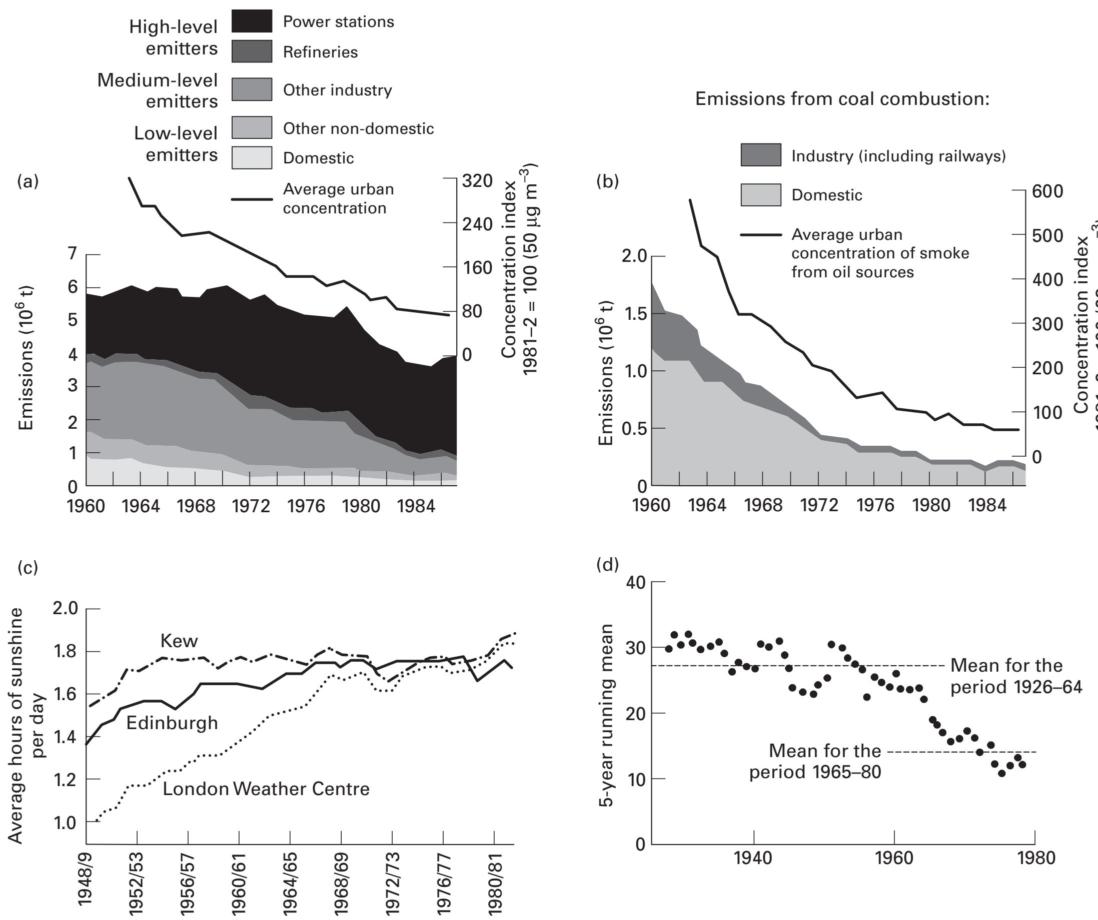

Figure 7.13 Trends in atmospheric quality in the UK: (a) sulfur dioxide emissions from coal combustion and average urban concentrations; (b) smoke emissions from coal combustion and average urban concentrations of oil smoke; (c) increase in winter sunshine (10-year moving average) for London and Edinburgh city centers and for Kew, outer London; (d) annual fog frequency at 0900 GMT in Oxford, central England, 1926–80 (after Department of the Environment data, and Gomez and Smith, 1984, figure 3).

Fenger (1999) has argued that the development of urban air pollution shows certain general historical trends (see Figure 7.14a). At the earlier stages air pollution increases to high levels. There then follows various abatement measures that cause a stabilization of air quality to occur. Levels of pollution then fall as high technology solutions are applied, although this may be countered to a certain extent by growth in vehicular traffic.

In some developed cities concentrations of pollutants have indeed tended to fall over recent decades. This can result from changes in industrial technology or from legislative changes (e.g., clean air legislation, restriction on car use, etc.). In many British cities, for example, legislation since the 1950s has reduced the burning of coal. As a consequence, fogs have become less frequent and the amount of sunshine has increased. Figure 7.13 shows the overall trends for the UK, and highlights the decreasing fog frequency and increasing sunshine levels (Musk, 1991). Similarly lead concentrations in the air in British cities have declined sharply following the introduction of unleaded petrol (Figure 7.14b) (Kirby, 1995), as they have in Copenhagen (Figure 7.14c) (Fenger, 1999). The concentrations of various pollutants have also been reduced in the Los Angeles area of California. Here, carbon monoxide, nonmethane hydrocarbon, nitrogen oxide, and ozone concentrations have all fallen steadily over the period since the late 1960s (Lents and Kelly, 1993).

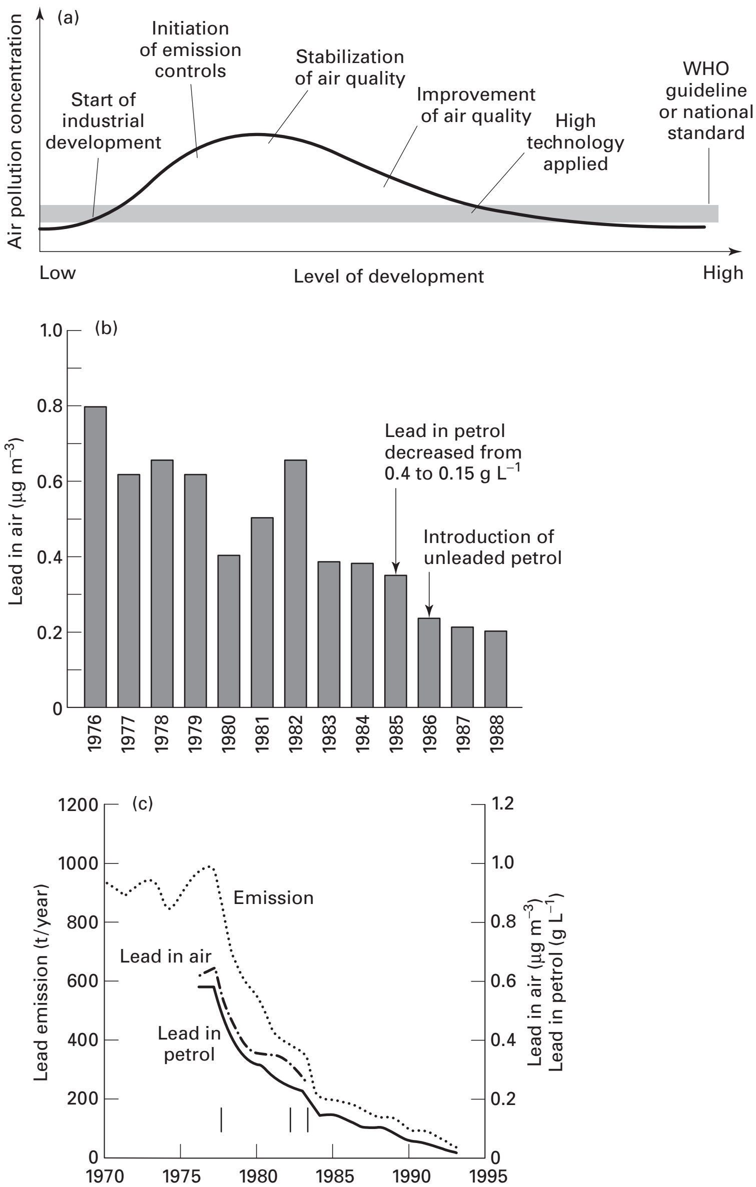

Figure 7.14 (a) Schematic presentation of a typical development of urban air pollution levels (after Fenger, 1999, figure 3): WHO – World Health Organization. (b) Lead concentration (annual means) in air at UK sites (Kirby, 1995, figure 1). (c) Annual average values for the total Danish lead emissions 1969–1993, the lead pollution in Copenhagen since 1976, and the average lead content in petrol sold in Denmark (Jensen and Fenger, 1994). The dates of tightening of restrictions on lead content are indicated with bars. Lead concentrations for the recent years can be found in Kemp et al. (1998). (Source: Fenger, 1999, figure 12.)

However, these three examples of improving trends come from developed countries. In many cities in poorer countries, pollution is increasing at present. In certain countries, heavy reliance on coal, oil, and even wood for domestic cooking and heating means that their levels of sulfur dioxide and suspended particulate matter (SPM) are high and climbing. In addition, rapid economic development is bringing increased emissions from industry and motor vehicles, which are generating progressively more serious air-quality problems.

Particular attention is being paid at the present time to the chemical composition of SPMs, and particularly to those particles that are small enough to be breathed in (i.e., smaller than 10 μm, and so often known as PM10s). Also of great concern in terms of human health are elemental carbon (e.g., from diesel vehicles), polycyclic aromatic hydrocarbons (PAHs), and toxic base metals (e.g., arsenic, lead, cadmium, and mercury), in part because of their possible role as carcinogens.

A major cause of urban air pollution is the development of photochemical smog. The name originates from the fact that most of the less desirable properties of such fog result from the products of chemical reactions induced by sunlight. Unburned hydrocarbons

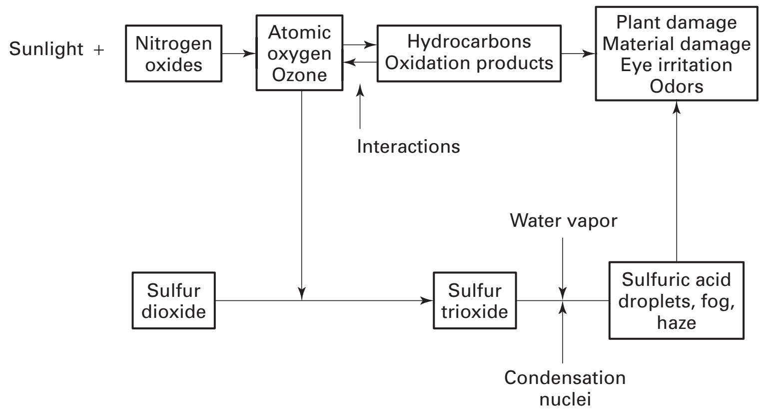

Figure 7.15 Possible reactions involving primary and secondary pollutants (after Haagen-Smit, in Bryson and Kutzbach, 1968, figure 4. Reprinted by permission of the Association of American Geographers).

play a major role in this type of smog formation and result from evaporation of solvents and fuels, as well as incomplete combustion of fossil fuels. In the presence of oxides of nitrogen, strong sunlight, and stable meteorological conditions, complex chemical reactions occur, forming a family of peroxyacyl nitrates (sometimes collectively abbreviated to PANs).

Photochemical smog appears ‘cleaner’ than other kinds of fog in the sense that it does not contain the very large particles of soot that are so characteristic of smog derived from coal burning. However, the eye irritation and damage to plant leaves it causes make it unpleasant. Photochemical smog occurs particularly where there is large-scale combustion of petroleum products, as in car-dominated cities such as Los Angeles. Its unpleasant properties include a high lead content; also a series of chemical reactions are triggered by sunlight (Figure 7.15). For example, a photochemical decomposition of nitrogen dioxide into nitric oxide and atomic oxygen occurs, and the atomic oxygen can react with molecular oxygen to form ozone. Further ozone may be produced by the reaction of atomic oxygen with various hydrocarbons.

Photochemical smogs are not universal. Because sunlight is a crucial factor in their development they are most common in the tropics or during seasons of strong sunshine. Their especial notoriety in Los Angeles is due to a meteorological setting dominated at times by subtropical anticyclones with weak winds, clear skies, and a subsidence inversion, combined with the general topographic situation and the high vehicle density (> 1500 vehicles per square kilometer). However, photochemical ozone pollution can, on certain summer days, with anticyclonic conditions bringing air in from Europe, reach appreciable levels even in the UK (Jenkins et al., 2002), especially in large cities such as London. Rigorous controls on vehicle emissions can greatly reduce the problem of high urban ozone concentrations and this has been a major cause of the reduction in ozone levels in Los Angeles over the past two decades, in spite of a growth in that city’s population and vehicle numbers (Figure 7.16). A full discussion of tropospheric ozone trends on a global basis is provided by Guicherit and Roemer (2000).

Air pollution: some further effects

This chapter has already made much reference to the ways in which humans have changed the turbidity of the atmosphere and the gases within it. However, the consequences of air pollution go further than either their direct impact on human health or their impact on local, regional, and global climates.

First of all, the atmosphere acts as a major channel for the transfer of pollutants from one place to another, so that some harmful substances have been transferred long distances from their sources of emission. Dichlorodiphenyltrichloroethane (DDT) is one example; lead is another. Thus, from the start of the industrial revolution, the lead content of the Greenland ice-cap, although far removed from the source of the pollutant (which is largely derived from either industrial or automobile emissions), rose very substantially (Figure 7.17a). The same applies to its sulfate content (Figure 7.17b). An analysis of pond sediments from a remote part of North America (Yosemite) indicates that lead levels were raised as a result of human activities,

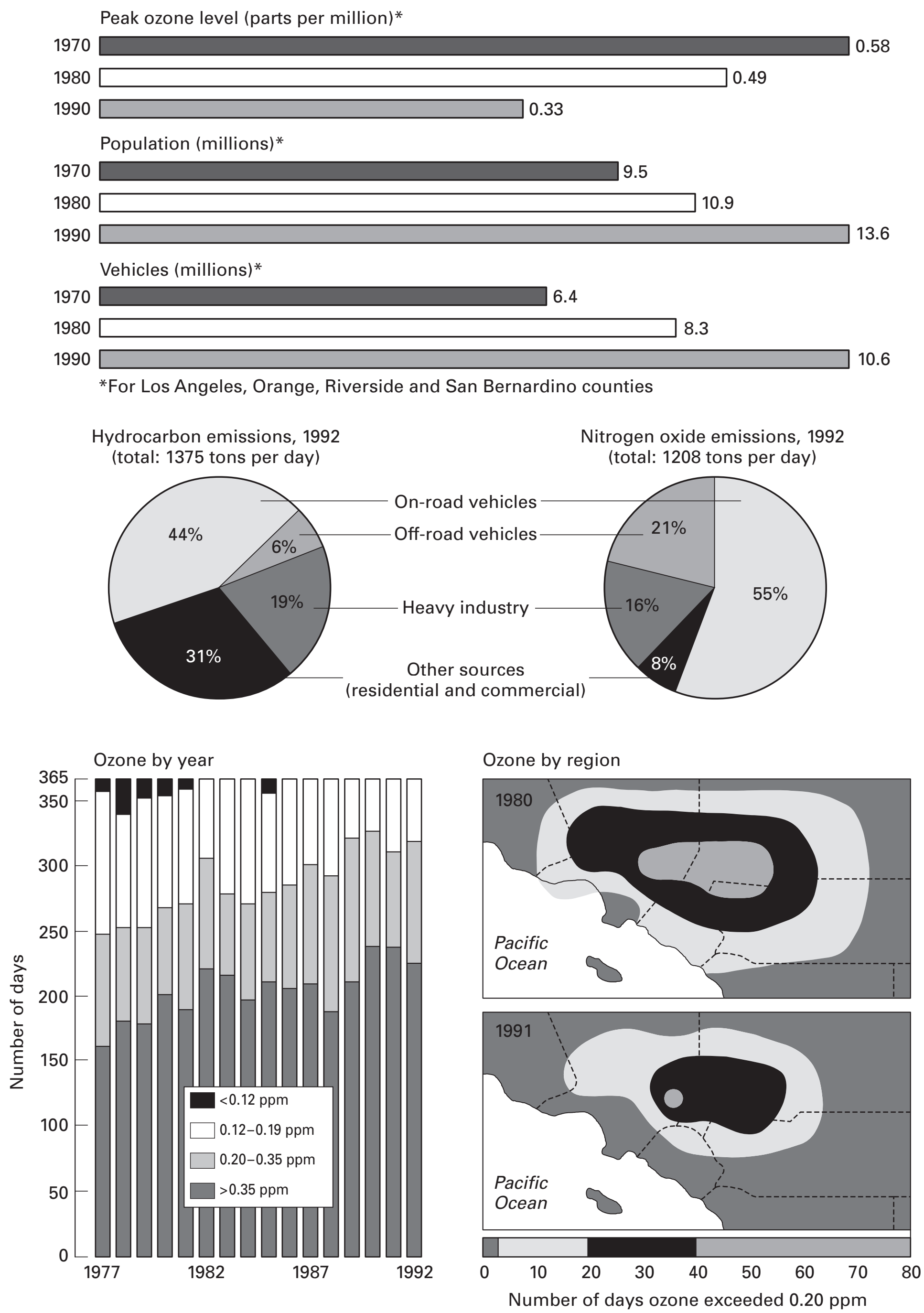

Figure 7.16 Air pollution in the Los Angeles area, 1970s to 1990s (after Lents and Kelly, 1993, p. 22).

being more than 20 times the natural levels. The lead, which came in from atmospheric sources, showed a fivefold elevation in the plants of the area and a fiftyfold elevation in the animals compared with natural levels (Shirahata et al., 1980). Some estimates also compared the total quantities of heavy metals that humans are releasing into the atmosphere with emissions from natural sources (Nriagu, 1979). The increase was eighteenfold for lead, ninefold for cadmium, sevenfold for zinc, and threefold for copper.

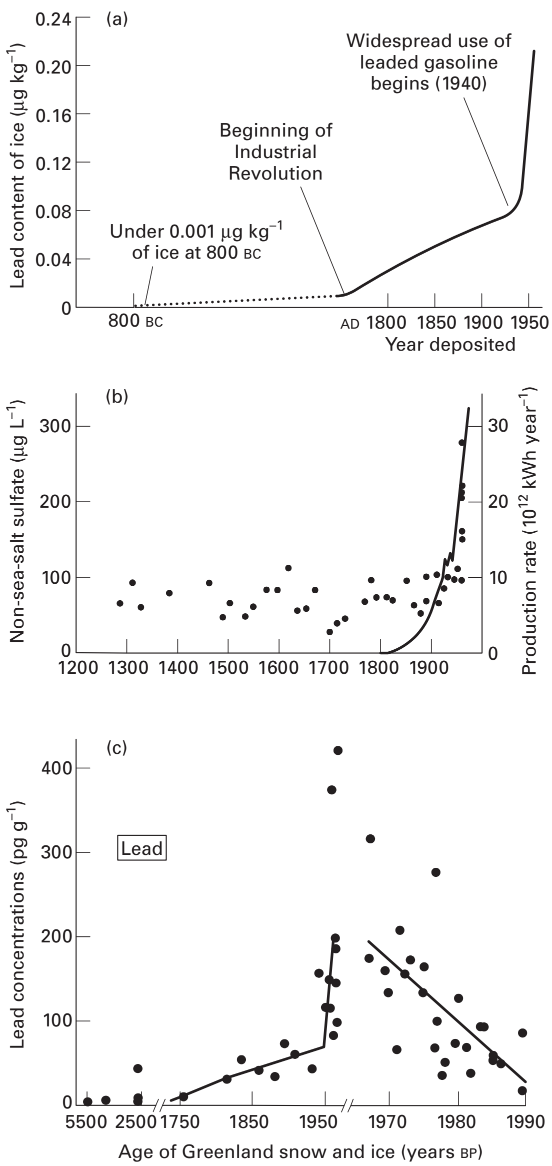

Figure 7.17 Trends in atmospheric quality. (a) Lead content of the Greenland ice-cap due to atmospheric fallout of the mineral on the snow surface. A dramatic upturn in worldwide atmospheric levels of lead occurred at the beginning of the industrial revolution in the nineteenth century and again after the more recent spread of the automobile (after Murozumi et al., 1969, p. 1247). (b) The sulfate concentration on a sea-salt-free basis in northwest Greenland glacier ice samples as a function of year. The curve represents the world production of thermal energy from coal, lignite, and crude oil (modified after Koide and Goldberg, 1971, figure 1). (c) Lead concentrations in Greenland snow (after Boutron et al., 1991. Reprinted with permission from *Nature*. Copyright 1991. Macmillan Magazines Limited).

There are, it must be stated, signs that as a result of pollution control regulations some of these trends are now being reversed. Boutron et al. (1991), for example, analyzed ice and snow that has accumulated over Greenland in the previous two decades and found that lead concentrations had decreased by a factor of 7.5 since 1970 (Figure 7.17c). They attribute this to a curbing of the use of lead additives in petrol. Over the same period cadmium and zinc concentrations have decreased by a factor of 2.5.

A second example of the possible widespread and ramifying ecological consequences of atmospheric pollution is provided by 'acid rain' (Likens and Bormann, 1974). Acid rain is rain which has a pH of less than 5.65, this being the pH which is produced by carbonic acid in equilibrium with atmospheric CO_2. In many parts of the world, rain may be markedly more acid than this normal, natural background level. Snow and rain in the northeast USA have been known to have pH values as low as 2.1, while in Scotland in one storm the rain was the acidic equivalent of vinegar (pH 2.4). In the eastern USA the average annual precipitation acidity values tend to be between pH 4 and 4.5 (see Figure 7.18), and the degree of acidulation appears to have increased between the 1950s and 1970s.

It needs to be remembered that not all environmental acidification is caused by acid rain in the narrow sense. Acidity can reach the ground surface without the assistance of water droplets. This is as particulate matter and is termed 'dry deposition'. Furthermore, there are various types of 'wet precipitation' by mist, hail, sleet, or snow, in addition to rain itself. Thus some people prefer the term 'acid deposition' to 'acid rain'. The acidity of the precipitation in turn leads to greater acidity in rivers and lakes.

The causes of acid deposition are the quantities of sulfur oxides (Figure 7.19) and nitrogen oxides emitted from fossil-fuel combustion. Figure 7.20 shows how sulfate levels increased in European precipitation between the 1950s and 1970s. Two main factors contributed to the increasing seriousness of the problem at the time. One was the replacement of coal by oil and natural gas. The second, paradoxically, was a result of the implementation of air pollution control measures (particularly increasing the height of smokestacks and installing particle precipitators). These appear to have transformed a local 'soot problem' into a regional 'acid rain problem'. Coal burning produced a great deal of

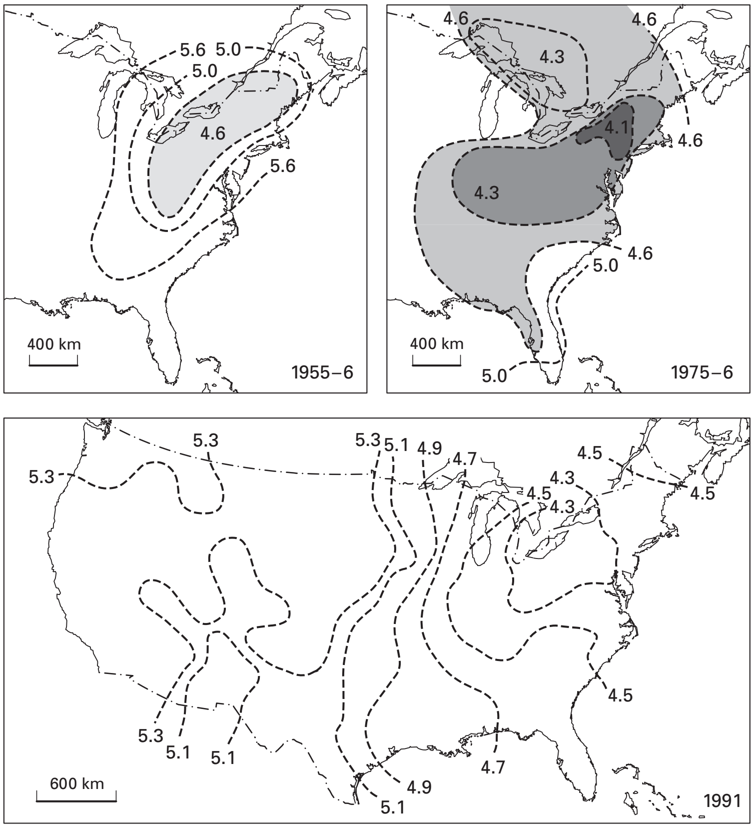

Figure 7.18 Isopleths showing average pH for precipitation in North America. Note the low values for eastern North America and the relatively higher values to the west of the Mississippi. (Combined from Likens et al., 1979 and Graedel and Crutzen, 1995, with modifications.)

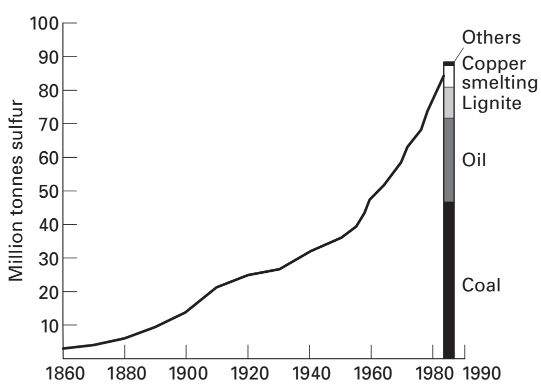

Figure 7.19 Global SO_2 emissions from anthropogenic sources, including the burning of coal, lignite, and oil, and copper smelting.

sulfate, but was largely neutralized by high calcium contents in the relatively unfiltered coal smoke emissions. Natural gas burning creates less sulfate but that which is produced is not neutralized. The new higher chimneys pump the smoke so high that it is dispersed over wide areas, whereas previously it returned to earth nearer the source.

Very long-term records of lake acidification provided dramatic evidence for the recent magnification of the acid deposition problem. These were obtained by extracting cores from lake floors and analyzing their diatom assemblages at different levels. The diatom assemblages reflect water acidity levels at the time they were living. Two studies serve to illustrate the trend. In southwest Sweden (Renberg and Hellberg, 1982) for most of post-glacial time (that is, the past 12,500 years) the pH of the lakes appears to have decreased gradually from around 7.0 to about 6.0 as a result of natural aging processes. However, especially since the 1950s, a marked decrease occurred to present-day values of about 4.5. In Britain, the work of Battarbee and collaborators (1985a and b) shows that at sensitive sites pH values before around 1850 were close to 6.0 and that since then pH declines have varied between 0.5 and 1.5 units. More monitoring of lake acidity also demonstrated that significant changes were taking place. Studies by Beamish et al. (1975)

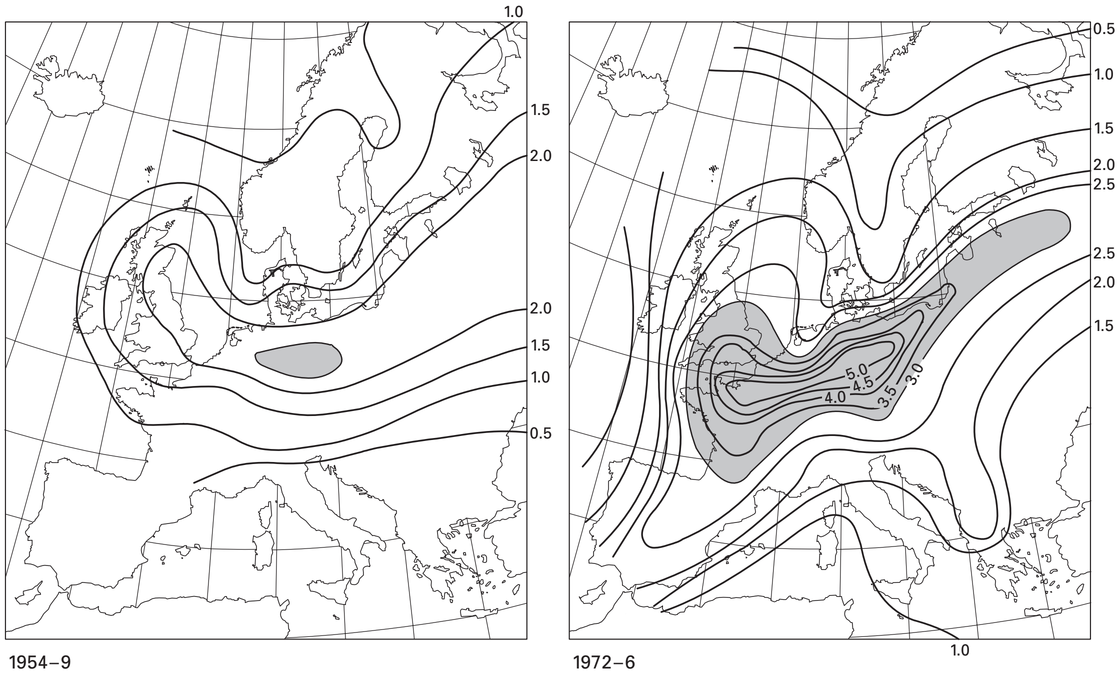

Figure 7.20 Annual mean concentration of sulfate in precipitation in Europe (mg S L^{-1}) (after Wallen in Holdgate et al., 1982, figure 2.3).

demonstrated that between 1961 and 1975 pH had declined by 0.13 pH units per year in George Lake, Canada, and a comparable picture emerges from Sweden, where Almer et al. (1974) found that the pH in some lakes had decreased by as much as 1.8 pH units since the 1930s.

The effects of acid rain (Figure 7.21) are especially serious in areas underlain by highly siliceous types of bedrock (e.g., granite, some gneisses, quartzite, and quartz sandstone), such as the old shield areas of the Fenno-Scandian shield in Scandinavia and the Laurentide shield in Canada (Likens et al., 1979). This is because of the lack of buffering by cations. Mobile anions are able to cause the leaching of basic cations (nutrients).

The ecological consequences of acid rain are still the subject of some debate. Krug and Frink (1983) have argued that acid rain only accelerates natural processes, and point out that the results of natural soil formation in humid climates include the leaching of nutrients, the release of aluminum ions and the acidification of soil and water. They also note that acidification by acid rain may be superimposed on longer-term acidification induced by changes in land use. Thus the regrowth of coniferous forests in what are now marginal agriculture areas, such as New England and highland western Europe, can increase acidification of soils and water. Similarly, it is possible, though in general unproven (see Battarbee et al., 1985a), that in areas such as western Scotland a decline in upland agriculture and the regeneration of heathland could play a role in increasing soil and water acidification. Likewise Johnston et al. (1982) have suggested that acid rain can cause either a decrease or an increase in forest productivity, depending on local factors. For example in soils where cation nutrients are abundant and sulfur or nitrogen are deficient, moderate inputs of acid rain are very likely to stimulate forest growth.

In general, however, it is the negative consequences of acid rain that have been stressed. One harmful effect is a change in soil character. The high concentration of hydrogen ions in acid rain causes accelerated leaching of essential nutrients, making them less available for plant use. Furthermore, the solubility of aluminum and heavy metal ions increases and instead of being fixed in the soil's sorption complex these toxic substances become available for plants or are transferred into lakes, where they become a major physiological stress for some aquatic organisms.

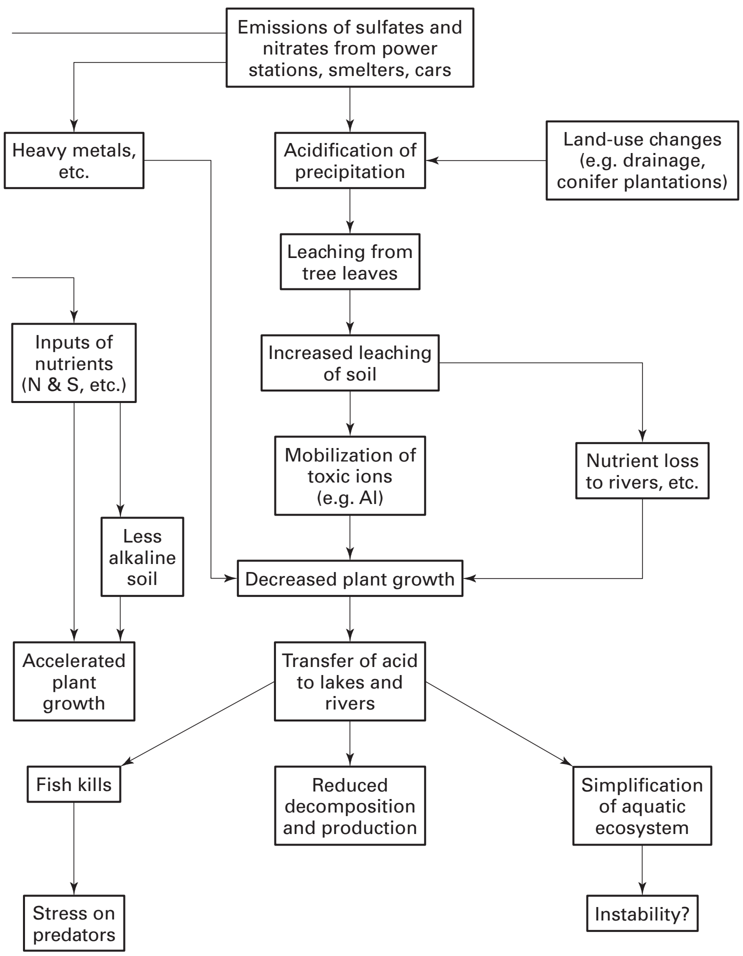

Figure 7.21 Pathways and effects of acid precipitation through different components of the ecosystem, showing some of the adverse and beneficial consequences.

Freshwater bodies with limited natural cations are poorly buffered and thus vulnerable to acid inputs. The acidification of thousands of lakes and rivers in southern Norway and Sweden during the late twentieth century has been attributed to acid rain, and this increased acidity resulted in the decline of various species of fish, particularly trout and salmon. But fish are not the only aquatic organisms that may be affected. Fungi and moss may proliferate, organic matter may start to decompose less rapidly, and the number of green algae may be reduced. Forest growth can also be affected by acid rain, although the evidence is not necessarily proven. Acid rain can damage foliage, increase susceptibility to pathogens, affect germination and reduce nutrient availability. However, since acid precipitation is only one of many environmental stresses, its impact may enhance, be enhanced by, or be swamped by other factors. For example, Blank (1985), in considering the fact that an estimated one-half of the total forest area of the former West Germany was showing signs of damage, referred to the possible role of ozone or of a run of hot, dry summers on tree health and growth. The whole question of forest decline is addressed in Chapter 4.

The seriousness of acid rain caused by sulfur dioxide emissions in the Western industrialized nations peaked in the mid-1970s or early 1980s (Figure 7.22). Changes in industrial technology, in the nature of economic activity, and in legislation caused the output of SO_2 in Britain to decrease by 35% between 1974 and 1990. This was also the case in many industrialized countries (Table 7.10b), including the USA (Malm et al., 2002). However, there has been a shift in the geographical sources of sulfate emissions, so that whereas in 1980 60% of global emissions were from the USA, Canada and Europe, by 1995 only 38% of world emis-

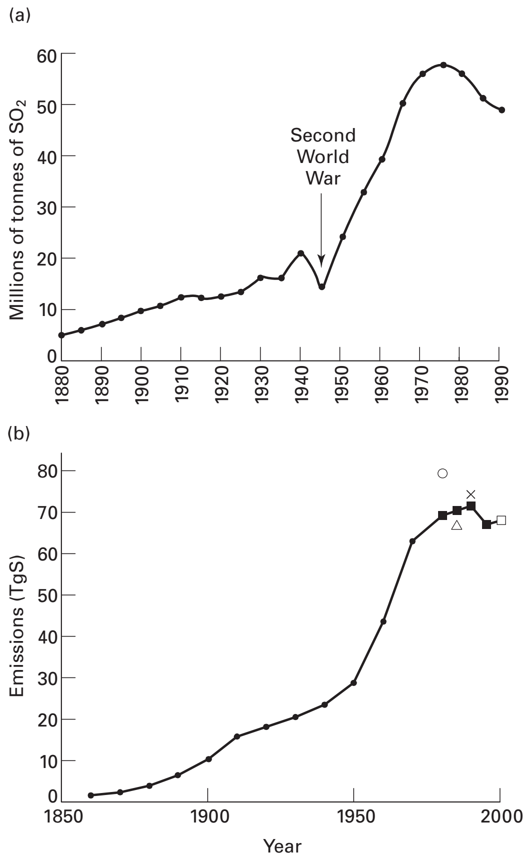

Figure 7.22 (a) Trends in sulfate emissions in Europe 1880–1990, based on data in Mylona (1996). (b) Estimates of historical total global sulfur dioxide (TgS) emissions from anthropogenic sources. From Smith et al. (2001, figure 1) with modifications.

sions originated from this region (Smith et al., 2001). There are also increasing controls on the emission of NO_x in vehicle exhaust emissions, although emissions of nitrogen oxides in Europe have not declined (Table 7.10). A similar picture emerges from Japan (Seto et al., 2002) where sulfate emissions have fallen because of emission controls, whereas nitrate emissions have increased with an increase in vehicular traffic. This means that the geography of acid rain may change, with it becoming less serious in the developed world, but with it increasing in locations such as China, where economic development will continue to be fueled by the burning of low quality sulfur-rich coal in enormous quantities.

In those countries where acid rain remains or is increasing as a problem there are various methods available to reduce its damaging effects. One of these is to add powdered limestone to lakes to increase their pH values. However, the only really effective and practical long-term treatment is to curb the emission of the offending gases. This can be achieved in a variety of ways: by reducing the amount of fossil fuel combustion; by using less sulfur-rich fossil fuels; by using alternative energy sources that do not produce nitrate or sulfate gases (e.g., hydropower or nuclear power); and by removing the pollutants before they reach the atmosphere. For example, after combustion at a power station, sulfur can be removed ('scrubbed') from flue gases by a process known as flue gas desulfurization (FGD), in which a mixture of limestone and water is sprayed into the flue gas which converts the sulfur dioxide (SO_2) into gypsum (calcium sulfate). Reducing NO_x in flue gas can be achieved by adding ammonia and passing it over a catalyst to produce nitrogen and water (a process called selective catalytic reduction or SCR), and NO_x produced by cars can be reduced by fitting a catalytic converter.

Stratospheric ozone depletion

A very rapidly developing area of concern in pollution studies is the current status of stratospheric ozone levels. The atmosphere has a layer of relatively high concentration of ozone (O_3) at a height of about 16 and 18 km in the polar latitudes and of about 25 km in equatorial regions. This ozone layer is important because it absorbs incoming solar ultraviolet radiation, thus warming the stratosphere and creating a steep inversion of temperature at heights between about 15 and 50 km. This in turn affects convective processes and atmospheric circulation, thereby influencing global weather and climate. However, the role of ozone in controlling receipts of ultraviolet radiation at the surface of the Earth has great ecological significance, because it modifies rates of photosynthesis. One class of organism that has been identified as being especially prone to the effects of increased ultraviolet radiation consequent upon ozone depletion is phytoplankton – aquatic plants that spend much of their time near the sea surface and are therefore exposed to such radiation. A reduction in their productivities would have potentially ramifying consequences, because these plants directly and indirectly provide the food for almost all

Table 7.10 Emissions of sulfur and nitrogen oxides (as NO_2). Source: Smith et al. (2001)
(a) For selected European countries in 1980 and 1993 (1000 t $year^{-1}$)

Country	*Sulfur (1000 t year^{-1})*			*Nitrogen oxides (1000 t year^{-1})*		
	1980	*1993*	*1993 as % of 1980*	*1980*	*1993*	*1993 as % of 1980*
Czech Republic	1128	710	62.9	937	574	61.26
Denmark	226	78	34.51	274	264	96.35
Finland	292	60	20.55	264	253	95.83
France	1669	568	34.03	1823	1519	83.32
Germany	3743	1948	52.04	2440	2904	84.42
Ireland	111	78	70.27	73	122	167.12
Italy	1900	1126	59.26	1480	2053	138.72
Netherlands	244	84	34.43	582	561	96.39
Norway	70	18	25.71	185	225	120.96
Poland	2050	1362	66.44	1500	1140	76.00
Spain	1660	1158	69.76	950	1257	132.32
Sweden	254	50	19.69	424	399	94.10
UK	2454	1597	65.08	2395	2355	98.32
Overall			47.28			103.47

(b) Percentage contributions for four regions to global sulfur dioxide emissions

Region	*Sulfur dioxide emissions by geographic region (%)*				
	1980	*1985*	*1990*	*1995*	*2000*
USA/Canada	21	18	18	16	15
Europe	39	36	31	22	19
Asia	26	31	35	43	46
Rest of world	14	15	16	18	20

fish. Ozone also protects humans from adverse effects of ultraviolet radiation, which include damage to the eyes, suppression of the immune system and higher rates of skin cancer.

Human activities appear to be causing ozone depletion in the stratosphere, most notably over the south polar regions, where an ‘ozone hole’ has been identified (Staehelin et al., 2001). Possible causes of ozone depletion are legion, and include various combustion products emitted from high-flying military and civil supersonic aircraft; nitrous oxide released from nitrogenous chemical fertilizers; and chlorofluorocarbons (CFCs) used in aerosol spray cans, refrigerant systems, and in the manufacture of foam fast-food containers.

However, in recent years the greatest attention has been focused on the role of CFCs, the production of which climbed greatly in the decade after the Second World War (Figure 7.23). These gases may diffuse upwards into the stratosphere where solar radiation causes them to become dissociated to yield chlorine atoms which react with and destroy the ozone. The process has been described thus by Titus and Seidel (1986: 4):

> Because CFCs are very stable compounds, they do not break up in the lower atmosphere (known as the troposphere). Instead, they slowly migrate to the stratosphere, where ultraviolet radiation breaks them down, releasing chlorine.
>
> Chlorine acts as a catalyst to destroy ozone; it promotes reactions that destroy ozone without being consumed. A chlorine (Cl) atom reacts with ozone (O_3) to form ClO and O_2. The ClO later reacts with another O_3 to form two molecules of O_2, which releases the Cl atoms. Thus two molecules of ozone are converted to three molecules of ordinary oxygen, and the chlorine is once again free to start the process. A single chlorine atom can destroy thousands of ozone molecules. Eventually, it returns to the troposphere, where it is rained out as hydrochloric acid.

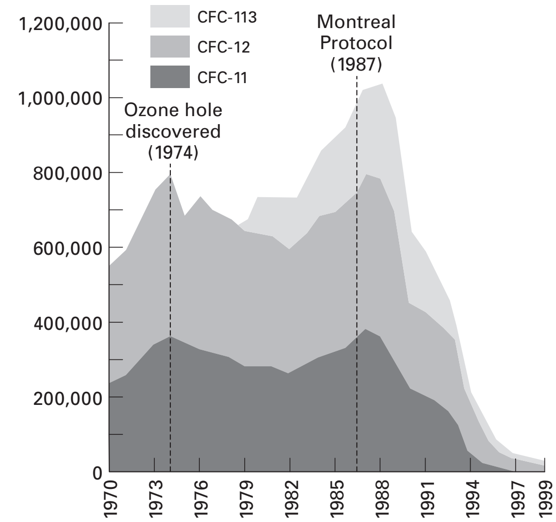

Figure 7.23 World production of major chlorofluorocarbons (t year^{-1}). World production of the three major CFCs peaked in about 1988 and has since declined to very low values.

The Antarctic ozone hole (Figure 7.24) has been identified through satellite monitoring and by monitoring of atmospheric chemistry on the ground. The decrease in ozone levels at Halley Bay is shown in Figure 7.25.

The reasons why this zone of ozone depletion is so well developed over Antarctica include: the very low temperatures of the polar winter, which seem to play a role in releasing chlorine atoms; the long sunlight hours of the polar summer, which promote photochemical processes; and the existence of a well-defined circulation vortex. This vortex is a region of very cold air surrounded by strong westerly winds, and air within this vortex is isolated from that at lower latitudes, permitting chemical reactions to be contained rather than more widely diffused. No such clearly defined vortex exists in the Northern Hemisphere, although ozone depletion does seem to have occurred in the Arctic as well (Proffitt et al., 1990). Furthermore, observations in the past few years indicate that the Antarctic ozone hole is spreading over wider areas and persisting longer into the Antarctic summer (Solomon, 1999). It is also possible that the situation could be worsened by emissions of volcanic ash into the atmosphere (as from Mount Pinatubo), for these can also cause chemical reactions that lead to ozone depletion (Mintzer and Miller, 1992).

Decreases in stratospheric ozone levels on a global basis have been analyzed by Harris et al. (2003). The most negative trends occur at mid- to high latitudes

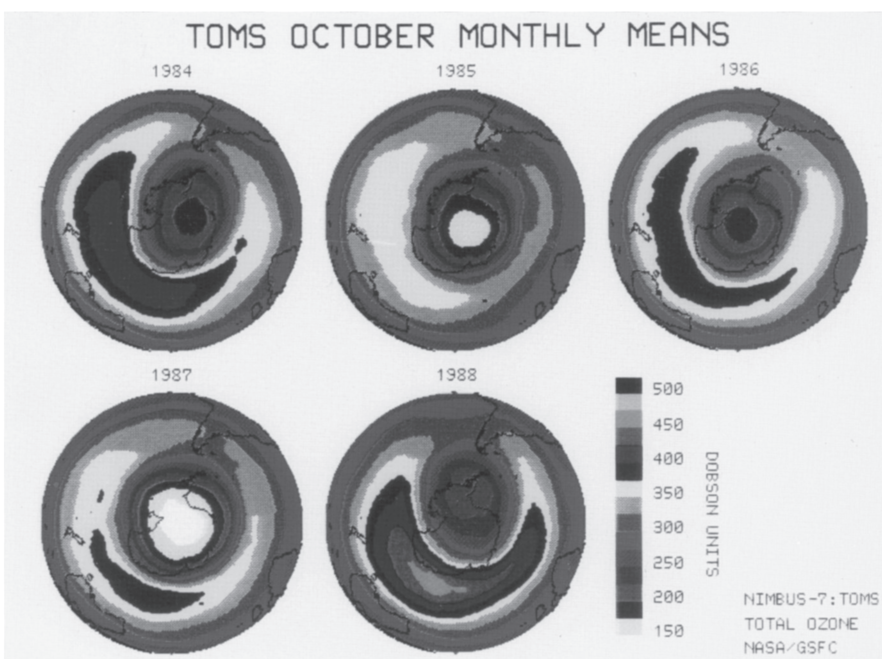

Figure 7.24 In the 1980s ground observations and satellite monitoring of atmospheric ozone levels indicated that a 'hole' had developed in the stratospheric ozone layer above Antarctica. These Nimbus satellite images show the ozone concentrations (in Dobson units) for the month of October between 1984 and 1988. Source: NASA.

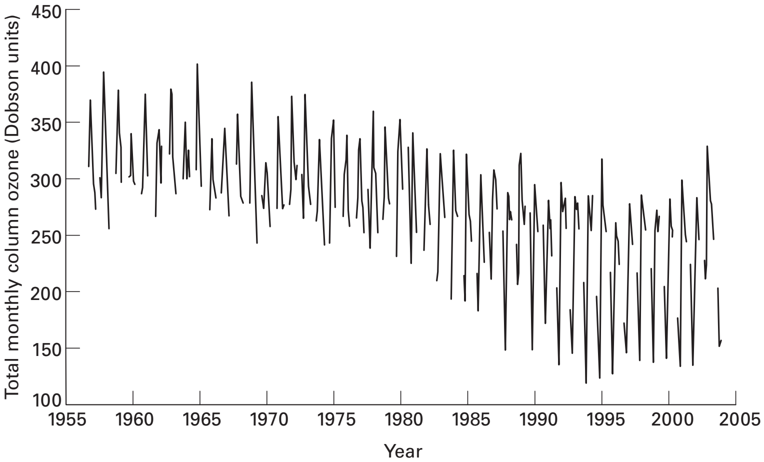

Figure 7.25 Total monthly ozone concentrations in Dobson units over Halley Bay, Antarctica (based on data provided by the British Antarctic Survey).

(Figure 7.26). Northern Hemisphere mid-latitude trends since 1978 have been at –2 to –3% per decade, while average losses in the Southern Hemisphere high latitude have been at up to 8% per decade.

Global production of CFC gases increased from around 180×10^6 kg per year in 1960 to nearly 1100×10^6 kg per year in 1990. However, in response to the thinning of the ozone layer, many governments signed an international agreement called the Montreal Protocol in 1987. This pledged them to a rapid phasing out of CFCs and halons. Production has since dropped substantially (Figure 7.23). However, because of their stability, these gases will persist in the atmosphere for decades or even centuries to come. Even with the most stringent controls that are now being considered, it will be the middle of the twenty-first century before the chlorine content of the stratosphere falls below the level that triggered the formation of the Antarctic 'ozone hole' in the first place.

Deliberate climatic modification

It has long been a human desire to modify weather and climate, but it is only since the Second World War and the development of high-altitude observations of clouds that serious attempts have been made to modify such phenomena as rainfall, hailstorms, and hurricanes (Cotton and Piehlke, 1995).

The most fruitful human attempts to augment natural precipitation have been through cloud seeding. Rainmaking experiments of this type are based on three main assumptions (Chorley and More, 1967: 159).

1 Either the presence of ice crystals in a supercooled cloud is necessary to release snow and rain, or the presence of comparatively large water droplets is necessary to initiate the coalescence process.
2 Some clouds precipitate inefficiently or not at all, because these components are naturally deficient.
3 The deficiency can be remedied by seeding the clouds artificially, either with solid carbon dioxide (dry ice) or silver iodide, to produce crystals, or by introducing water droplets or large hygroscopic nuclei (e.g., salt).

These methods of seeding are not universally productive and in many countries expenditure on such procedures has been reduced. Under conditions of orographic lift and in thunderstorm cells, when nuclei are insufficient to generate rain by natural means, some augmentation may be attained, especially if cloud temperatures are of the order of –10° to –15°C. The increase of precipitation gained under favorable conditions may be of the order of 10–20% in any one storm. In lower latitudes, where cloud-top temperatures frequently remain above 0°C, silver-iodide or dry-ice seeding is not applicable. Therefore alternative methods have been introduced whereby small water droplets of 50 mm diameter are sprayed into the lower layers of deep clouds, so that the growth of cloud particles will be stimulated by coalescence. Other techniques of warm cloud seeding include the feeding of hygroscopic

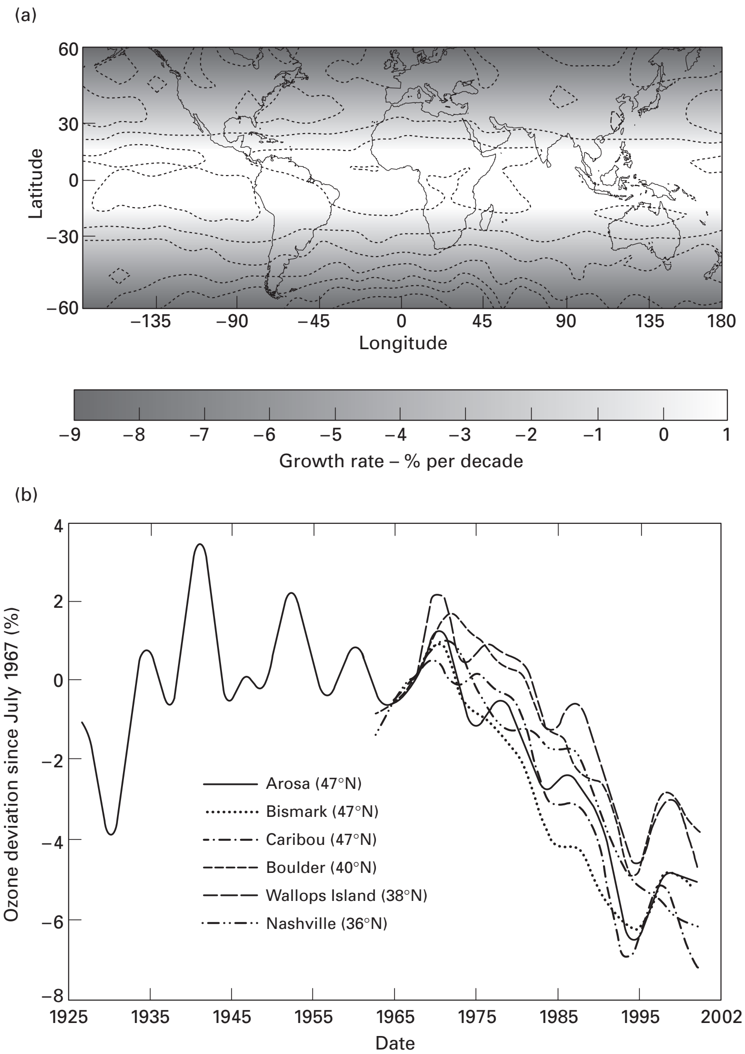

Figure 7.26 Global ozone trends: (a) average total ozone growth rates (percent per decade); (b) total ozone tendency curves for six northern mid-latitude sites. These sites are normalized to zero in July 1967, the beginning of the shortest record. Ozone deviation is given in percent. (Source: Harris et al., 2003, figures 4 and 7.)

particles into the lower air layers near the updraught of a growing cumulus cloud (Breuer, 1980).

The results of the many experiments now carried out on cloud seeding are still controversial, very largely because we have an imperfect understanding of the physical processes involved. This means that the evidence has to be evaluated on a statistical rather than a scientific basis so, although precipitation may occur after many seeding trials, it is difficult to decide to what extent artificial stimulation and augmentation is responsible. It also needs to be remembered that this form of planned weather modification applies to small areas for short periods. As yet, no means exist to change precipitation appreciably over large areas on a sustained basis.

Other types of deliberate climatic modification have also been attempted (Hess, 1974). For example, it has been thought that the production of many more hailstone embryos by silver-iodide seeding will yield smaller hailstones which would both be less damaging and more likely to melt before reaching the ground (Figure 7.27). Some results in Russia have been encouraging, but one cannot exclude the possibility that seeding may sometimes even increase hail damage (Atlas, 1977). Similar experiments have been conducted in lightning suppression. The concept here is to produce in a thundercloud, again by silver-iodide seeding, an abnormal abundance of ice crystals that would act as added corona points and thus relieve the electrical potential gradient by corona discharge before a

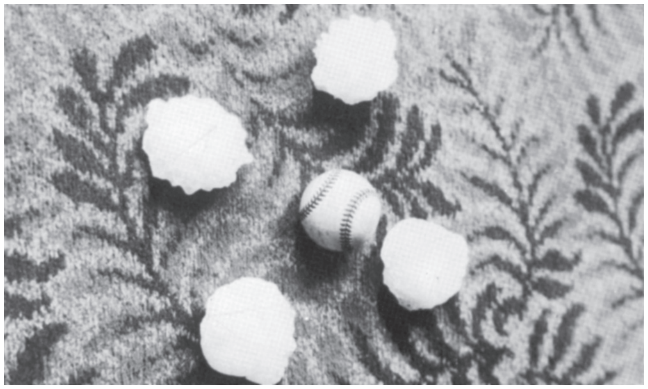

Figure 7.27 The damage caused by large hailstones can be considerable. These examples, the size of a baseball, fell near Neligh, Nebraska, in June 1950. There is, therefore, a considerable incentive to explore ways of reducing their impact through climatic modification experiments.

lightning strike could develop (Panel on Weather and Climate Modification, 1966: 4–8).

Hurricane modification is perhaps the most desirable aim of those seeking to suppress severe storms, because of the extremely favorable benefit-to-cost ratio of the work. The principle once again is that of introducing freezing nuclei into the ring of clouds around the hurricane center to trigger the release of the latent heat of fusion in the eye-wall cloud system which, in turn, diminishes the maximum horizontal temperature gradients in the storm, causing a hydrostatic lowering of the surface temperature. This eventually should lead to a weakening of the damaging winds (Smith, 1975: 212). A 15% reduction in maximum winds is theoretically possible, but as yet work is largely at an experimental stage. A 30% reduction in maximum winds was claimed following seeding of Hurricane Debbie in 1969 but attempts in the USA to seed hurricanes were discontinued in the 1970s following the development of new computer models of the effects of seeding on hurricanes. The models suggested that although maximum winds might be reduced by 10–15% on average, seeding may increase the winds just outside the region of maximum winds by 10–15%, may either increase or decrease the maximum storm surge, and may or may not affect the direction of the storm (Sorkin, 1982). Seeding of hurricanes is unlikely to recommence until such uncertainties can be resolved.

Much research has been devoted in Russia to the possibility of removing the Arctic Sea ice and to ascertain effects of such action on the climate of northern areas (see Lamb, 1977: 660). One proposal was to dam the Bering Strait, thereby blocking off water flow from the Pacific. The assumption was that more, and warmer, Atlantic water would be drawn into the central Arctic, improving temperature conditions in that area. Critics have pointed to the possibility of adverse changes in temperatures elsewhere, together with undesirable change in precipitation character and amount.

Fog dispersal, vital for airport operation, is another aim of weather modification. Seeding experiments have shown that fog consisting of supercooled droplets can be cleared by using liquid propane or dry ice. In very cold fogs this seeding method causes rapid transformation of water droplets into ice particles. Warm fogs with temperatures above freezing point occur more frequently than supercooled fogs in mid-latitudes and are more difficult to disperse. Some success has been achieved using sodium chloride and other hygroscopic particles as seeding agents but the most effective method is to evaporate the fog. The French have developed the 'turboclair' system in which jet engines are installed alongside the runway at major airports and the engines produce short bursts of heat to evaporate the fog and improve visibility as an aircraft approaches (Hess, 1974).

In regions of high temperature, dark soils become overheated, and the resultant high evapotranspiration rates lead to moisture deficiencies. Applications of white powders (India and Israel) or of aluminum foils (Hungary) increase the reflection from the soil surface and reduce the rate at which insolation is absorbed. Temperatures of the soil surface and subsurface are lowered (by as much as 10°C), and soil moisture is conserved (by as much as 50%).

The planting of windbreaks is an even more important attempt by humans to modify local climate deliberately. Shelterbelts have been in use for centuries in many windswept areas of the globe, both to protect soils from blowing and to protect the plants from the direct effects of high velocity winds. The size and effectiveness of the protection depends on their height, density, shape, and frequency. However, the belts may have consequences for microclimate beyond those for which they were planted. Evaporation rates are curtailed, snow is arrested, and its melting waters are available for the fields, but the temperatures may become more extreme in the stagnant space in the lee of the belt, creating an increase in frost risk.

Traditional farmers in many societies have been aware of the virtues of microclimatic management (Wilken, 1972). They manage shade by employing layered cropping systems or by covering the plant and soil with mulches; they may deliberately try to modify albedo conditions. Tibetan farmers, for example, reportedly throw dark rocks on to snow-covered fields to promote late spring melting; and in the Paris area of France some very dense stonewalls were constructed to absorb and radiate heat.

Conclusion

Changes in the composition of Earth's atmosphere as a result of human emissions of trace gases, and changes in the nature of land cover, have caused great concern in recent years. Global warming, ozone depletion and acid rain have become central issues in the study of environmental change. Although most attention is often paid to climatic change resulting from greenhouse gases, there is a whole series of other mechanisms that have the potential to cause climatic change. Most notably, this chapter has pointed to the importance of other changes in atmospheric composition and properties, whether these are caused by aerosol generation or albedo change.

However, the greenhouse effect and global warming may prove to have great significance for the environment and for human activities. Huge uncertainties remain about the speed, degree, direction, and spatial patterning of potential change. Nonetheless, if the earth warms up by a couple of degrees over the next 100 or so years, the impacts, some negative and some positive, are unlikely to be trivial. This is something that will form a focus of the following chapters.

For many people, especially in cities, the immediate climatic environment has already been changed. Urban climates are different in many ways from those of their rural surroundings. The quality of the air in many cities has been transformed by a range of pollutants, but under certain circumstances clean air legislation and other measures can cause rapid and often remarkable improvements in this area.

The same is true of two major pollution issues – ozone depletion and acid deposition. Both processes have serious environmental consequences and their effects may remain with us for many years, but both can be slowed down or even reversed by regulating the production and output of the offending gases.

Points for review

What do you understand by the term 'the greenhouse effect'?

Explain the major processes involved in global warming.

What role may changes in aerosols and albedo play in climate change?

How does the climate of cities differ from that of the surrounding countryside?

Give examples of some locations where levels of air pollution are falling. Why are they?

What is acid rain and how does it affect the environment?

What is 'the ozone hole' and how did it form?

Guide to reading

Elsma, D., 1996, *Smog alert: managing urban air quality*. London: Earthscan. A very readable and informative guide.

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Williams, M. A. J. and Balling, R. C., 1996, *Interactions of desertification and climate*. London: Arnold. A study of land cover changes and their effects.