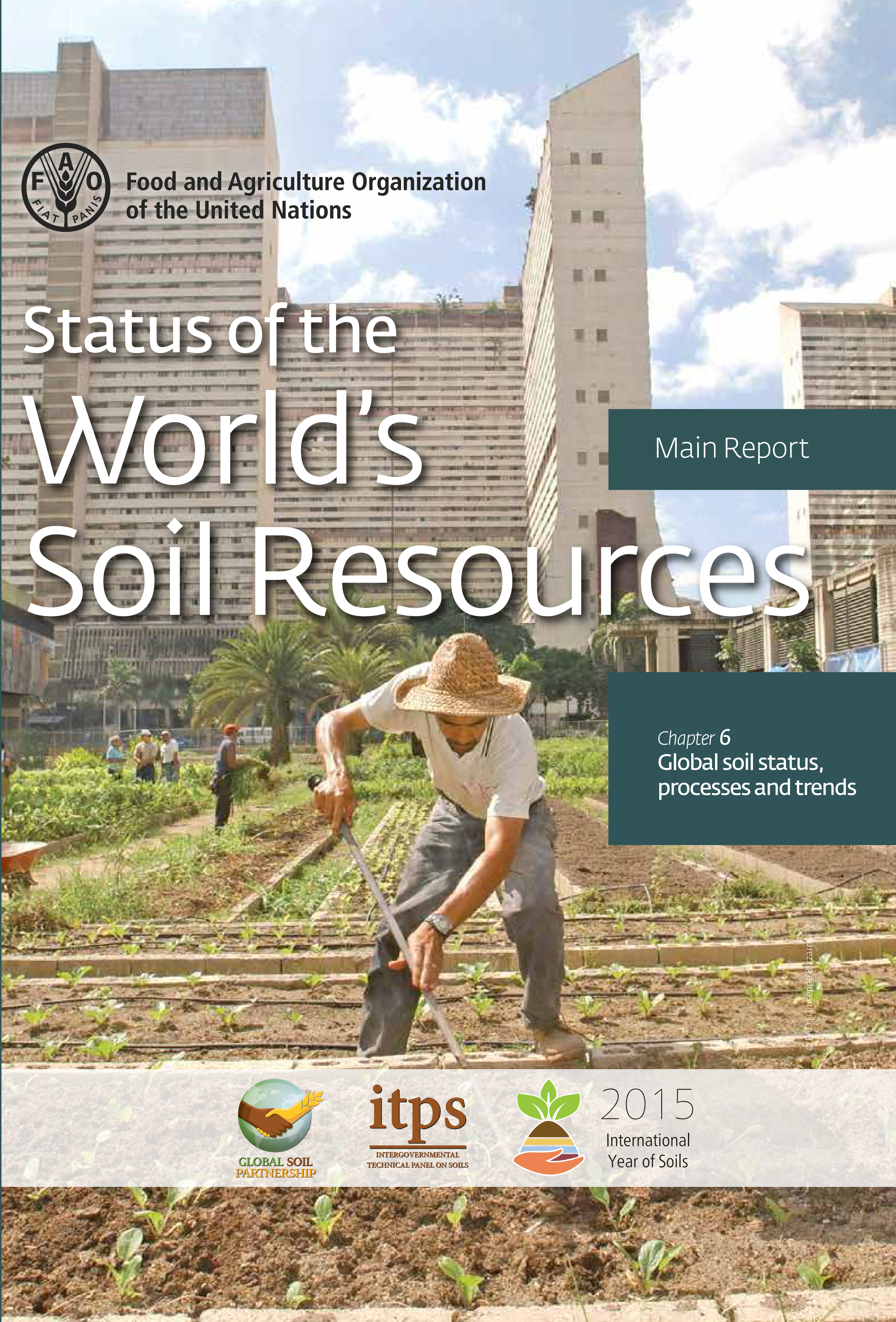

Food and Agriculture Organization of the United Nations

Status of the World's Soil Resources

Main Report

Chapter 6
Global soil status, processes and trends

GLOBAL SOIL PARTNERSHIP

itps
INTERGOVERNMENTAL TECHNICAL PANEL ON SOILS

2015 International Year of Soils

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6 | Global soil status, processes and trends

6.1 | Global status, processes and trends in soil erosion

6.1.1 | Processes

Soil erosion is broadly defined as the accelerated removal of topsoil from the land surface through water, wind or tillage. *Water erosion* on agricultural land occurs mainly when overland flow entrains soil particles detached by drop impact or runoff, often leading to clearly defined channels such as rills or gullies. *Wind erosion* occurs when dry, loose, bare soil is subjected to strong winds. Wind erosion is common in semi-arid areas where strong winds can easily mobilize soil particles, especially during dry spells. This dynamic physical aeolian process includes the detachment of particles from the soil, transport for varying distances depending on site, particle and wind characteristics, and subsequent deposition in a new location, causing onsite and offsite effects. During wind erosion events, larger particles creep along the ground or saltate (bounce) across the surface until they are deposited relatively close to field boundaries (Hagen *et al.*, 2007). Finer particles (< 80 µm) can travel great distances, with the finest particles entering global circulation (Shao, 2000). Tillage erosion is the direct down-slope movement of soil by tillage implements where particles only redistribute within a field.

6.1.2 | Status of Soil Erosion

Over the last decade, the figures published for *water erosion* range over an order of magnitude of ca. 20 Gt (gigaton) yr^{-1} to over 200 Gt yr^{-1}. While this huge variation may at first seem to suggest that our estimates of global soil erosion are very uncertain, a more detailed analysis shows that estimates exceeding ca. 50 Gt yr^{-1} are not realistic. In most cases, excessively high estimates can be traced back to conceptually flawed approaches and/or inappropriate model applications. Considering only those estimates that are not manifestly affected by such problems, the most likely range of global soil erosion by water is 20–30 Gt yr^{-1}, while *tillage erosion* may amount to ca. 5 Gt yr^{-1}.

Total erosion rates for *wind erosion* are highly uncertain. Estimates of the total amount of dust that is yearly mobilized on land place an upper limit on dust mobilization by wind erosion on arable land at ca. 2 Gt yr^{-1}. However, wind not only mobilizes dust but also coarser soil particles (sand), implying much higher total wind erosion rates. A large number of studies have made global estimates of wind erosion and dust transport. Approximately 430 million ha of drylands, which comprise 40 percent of the Earth's surface (Ravi *et al.*, 2011), are susceptible to wind erosion (Middleton and Thomas, 1997). In a survey of global estimates of present-climate dust emissions, Shao *et al.* (2011) described 13 studies that estimated global dust emissions in a range from 500 to ~ 3320 Tg yr^{-1}. Ginoux *et al.* (2012). The studies used global-scale high-resolution satellite imagery to study dust sources. They found that natural dust sources do account for about 75 percent of dust emissions and the remaining 25 percent of emissions were attributed to anthropogenic sources. The fraction of dust sources was highly variable. For example, although North Africa accounted for about 55 percent of the global dust emissions, only 8 percent originated from anthropogenic sources. In contrast, anthropogenic dust sources contributed 75 percent of the dust emissions in Australia (Ginoux *et al.*, 2012).

Translating these global estimates into accurate local soil erosion rates is not straightforward as soil erosion is highly variable, both in time and in space. However, typical soil erosion rates by water can be defined for representative agro-ecological conditions. Hilly croplands under conventional agriculture and orchards without additional soil cover in temperate climate zones are subject to erosion rates up to 10-20 tonnes ha^{-1} yr^{-1} , while average rates are often < 10 tonnes ha^{-1} yr^{-1}. Values during high-intensity rainfall events may reach 100 tonnes ha^{-1} and lead to muddy flooding in downstream areas. Erosion rates on hilly croplands in tropical and subtropical areas may reach values up to 50-100 tonnes ha^{-1} yr^{-1}. Average rates, however, are lower and often 10-20 tonnes $ha^{-1}yr^{-1}$. These high rates are due to the combination of an erosive climate (high intensity rainfall) and slope gradients which are generally steeper than those on cultivated land in the temperate zones. The incidence of erosion on steep slopes is due not only to specific topographic conditions, but also to the combination of a high population pressure with low-intensity agriculture, leading to the cultivation of marginal steeplands.

Rangelands and pasturelands in hilly tropical and sub-tropical areas may suffer erosion rates similar to those of tropical croplands. Due to the lack of field boundaries, which often act as barriers for sediment and runoff and promote infiltration, these rangelands may also be particularly vulnerable to gully formation. This may not affect topsoil so much but may make land inaccessible and hence unusable. Rangelands and pasturelands in temperate areas are characterized by erosion rates which are generally much lower and are most often below 1 tonnes ha^{-1} yr^{-1}. These rangelands are less intensively used and better managed than (sub-) tropical rangelands.

It is possible to identify the areas in the world where soil erosion by water is problematic based on a relatively simple modelling approach combining information on soil type, land use, topography and climate (Doetterl, Van Oost and Six, 2012; Van Oost *et al.*, 2007). Soil erosion by water is problematic in much of the hilly areas that are used as croplands on all continents, even where there have been significant conservation efforts as in the Mid-West of United States. Cropland in Europe is characterized by somewhat lower, yet still very significant soil erosion rates (Figure 6.1).

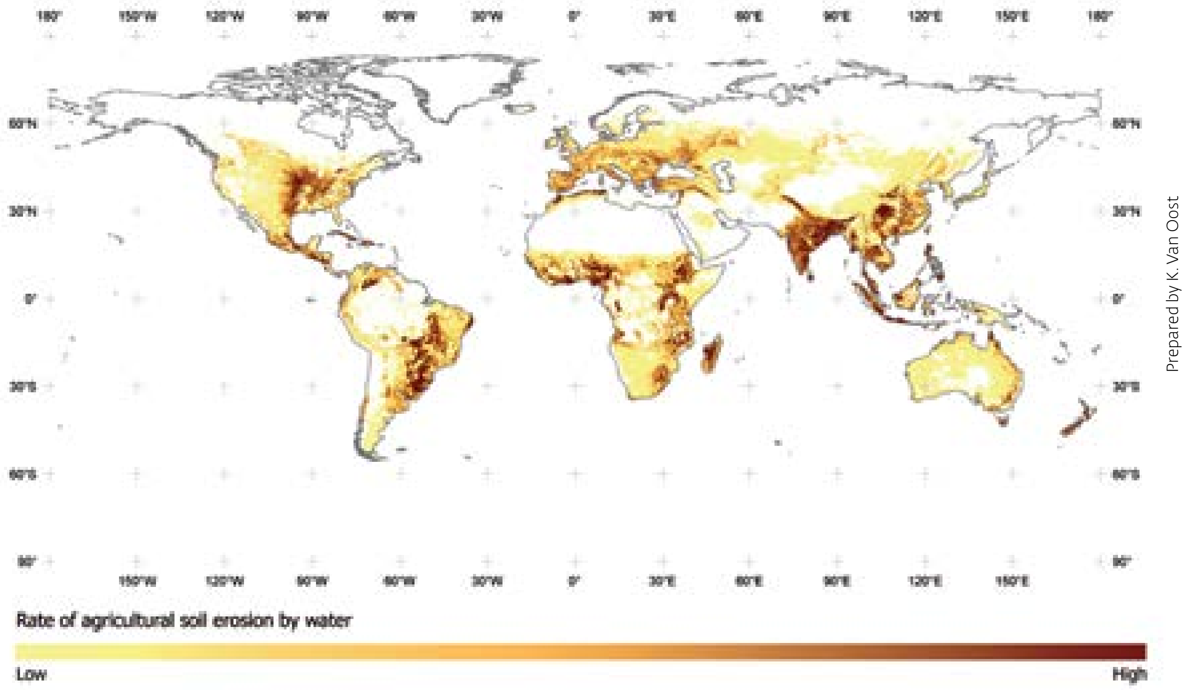

Figure 6.1 | Spatial variation of soil erosion by water. High rates (>ca. 20 t ha^{-1}y^{-1}) mainly occur on cropland in tropical areas.

The map gives an indication of current erosion rates and does not assess the degradation status of the soils. The map is derived from Van Oost et al., 2007 using a quantile classification.

The redistribution of soil within fields due to tillage erosion rates may lead to (very) high erosion rates on convexities (knolls) exceeding 30 tonnes ha^{-1} yr^{-1}; and to deposition rates in hollows and at down slope field borders exceeding 100 tonnes ha^{-1} yr^{-1}. These rates are not directly comparable to wind or water erosion rates, as soil eroded by tillage will not leave the field. However, tillage erosion may significantly reduce crop productivity on convexities and near upslope field or terrace borders.

Evidence of past erosion is extensive. This is demonstrated by wind-blown sands of sandstone bedrock, extensive loess accumulations of silt-sized aeolian sediments, and other formerly aeolian-affected landscapes. Large areas of sand seas, dune fields and other aeolian features and observations of activity provide further evidence of past wind erosion (Figure 6.2).

USDA estimates place wind erosion rates at ca. 2.5 tonnes ha^{-1} yr^{-1} on average over all cropland of the United States while the average erosion rate for pastureland is ca. 0.1 tonnes ha^{-1} yr^{-1}. There are very few quantitative assessments of wind erosion rates on arable land outside of the United States.

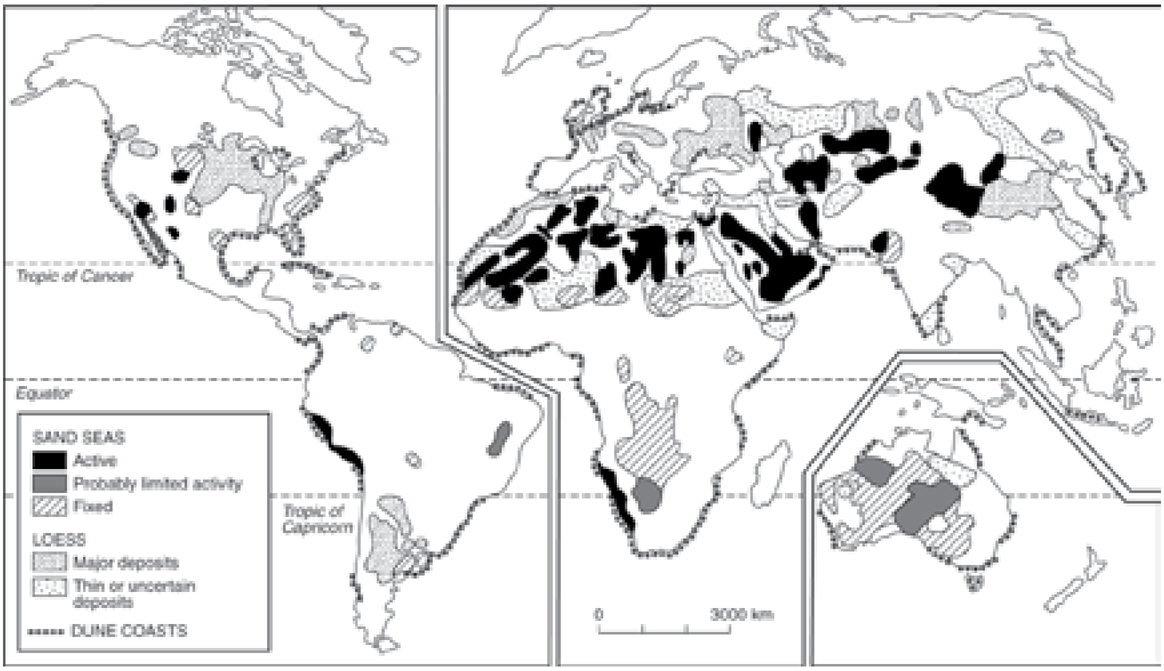

Figure 6.2 | Location of active and fixed aeolian deposits.
Source: Thomas and Wiggs, 2008.

6.1.3 | Soil erosion versus soil formation

The accelerated loss of topsoil through erosion from agricultural land was recognized as an important threat to the world's soil resource many decades ago. Furthermore, it was feared that soil was, in many areas, eroding much faster than that it could be replaced through soil formation processes. More recent studies have confirmed that these early observations were not just perceptions. Estimated rates of soil erosion of arable or intensively grazed lands have been found to be 100-1000 times higher than natural background erosion rates. These erosion rates are also much higher than known soil formation rates which are typically well below 1 tonnes ha^{-1} yr^{-1} with median values of ca. 0.15 tonnes ha^{-1} yr^{-1}. The large difference between erosion rates under conventional agriculture and soil formation rates implies that we are essentially mining the soil and that we should consider the resource as non-renewable.

The imbalance between erosion rates under conventional agriculture and the rate of soil formation implies that conventional agriculture on hilly land is not sustainable because the soil resource is mined and will ultimately become depleted. This has most likely already happened in many areas around the Mediterranean Sea and in tropical mountain regions. So-called soil loss tolerance levels may help to set objectives for short-term action. However, long-term sustainability requires that soil erosion rates on agricultural land are reduced to near-zero levels.

Figure 6.3 | Soil relict in the Jadan basin, Ecuador. Photo by G. Govers

In this area overgrazing led to excessive erosion and the soil has been completely stripped from most of the landscape in less than 200 years, exposing the highly weathered bedrock below. The person is standing on a small patch of the B-horizon of the original soil that has been preserved. Picture credit: Gerard Govers.

6.1.4 | Soil erodibility

Soil erodibility refers to the degree or intensity of a soil's state or susceptibility to being eroded (SSSA, 2008). A critical review of research into the factors controlling susceptibility of soils to wind erosion ('soil wind erodibility') has been provided by Webb and Strong (2011). Factors controlling soil wind erodibility include physical, chemical and biological characteristics of the soil, including texture, aggregation, stability, crusting, the amount of loose erodible sediment available, soil water content, roughness due to surface features (including tillage marks and vegetation) etc. The factors controlling soil wind erodibility differ somewhat among land uses and management approaches. For example, the factors controlling erodibility on rangeland differ from the factors controlling erodibility on farmland. In cropped soils at the field scale, disturbance due to tillage modifies the soil surface roughness, the amount and distribution of surface cover, soil water content aggregation and other properties, all of which affect soil erodibility for short periods of time (Zobeck, 1991; Zobeck and Van Pelt, 2011). In arid and semiarid rangeland ecosystems, wind erosion depends on vegetative cover and patchiness (Okin *et al.*, 2009) and on surface soil texture and crusting, characteristics that change more slowly unless disturbed. Not only do the factors controlling erosion by wind differ among land uses, but differences occur in their spatial and temporal variability. Natural and anthropogenic disturbances such as grazing, fire and other activities alter the surface and vegetation on rangeland while crop management practices often control erodibility of farmland. Webb and Strong (2011) described the dynamics of soil erodibility as a continuum that responds to changes in climate variability and disturbance. Factors such as rain and crusting on some soils may initially produce low erodibility that will subsequently increase with disturbance and drying. The exact timing and variability of changes in erodibility will vary with inherent soil physical properties such as soil texture.

6.1.5 | Soil erosion and agriculture

Soil erosion has direct, negative effects for global agriculture. Soil erosion by water induces annual fluxes of 23-42 Mt (megaton) N and 14.6-26.4 Mt P off agricultural land. These fluxes may be compared to annual fertilizer application rates, which are ca. 112 Tg for N and ca. 18 Tg of P. These nutrient losses need to be replaced through fertilization at a significant economic cost. Using a United States farm price of ca. US$ 1.45 per kg of N and ca. US$ 5.26 per kg of P implies an annual economic cost of US$ 33-60 billion for N and US$ 77-140 billion for P[1]. It is therefore clear that compensation for erosion-induced nutrient losses requires a massive investment in fertilizer use. In poor regions such as sub-Saharan Africa, the economic resources to achieve compensations for nutrient losses do not exist. As a consequence, the removal of nutrients by erosion from agricultural fields is much higher than the amount of fertilizer applied.

The detrimental removal of soil and nutrients from upland fields may be partly offset through the deposition of the eroded soil and nutrients in depositional areas. While this is true, such gains should not be exaggerated: the deposition of sediments and nutrients in large floodplains is not directly coupled to actual agricultural soil erosion, as in most cases sediments are provided by other sources (natural erosion, landslides) and the residence time of such sediments in large river systems is several thousands of years. In other words: the sediments that are currently being deposited in the Nile valley are not coming from the soils that are currently being eroded in Ethiopia. On a smaller scale, the deposition of eroded sediment may indeed locally increase local crop productivity, but such effects may be overshadowed by other factors, such as water availability.

Soil erosion does not induce an important carbon loss from the soil to the atmosphere: erosion mostly induces a transfer of carbon from eroding locations to depositional locations. Net losses are limited as the carbon lost at eroding locations is partially replaced through dynamic replacement whereas the soil carbon that is deposited in colluvial and alluvial settings may be stored there for several centuries.

1 www.ers.usda.gov/data-products/fertilizer-use-and-price.aspx#26727

High soil erosion rates will also have significant negative effects over longer time spans: the loss of topsoil will result in a reduction in the soil's capacity to provide rooting space and, more importantly, in the capacity to store water that can be released to plants. This may reduce soil productivity. However, these changes occur relatively slowly: the reduction in soil water holding capacity and and/or root space accommodation results in yield declines of ca. 4 percent per 0.1m of soil lost. Except for areas where erosion rates are very high (e.g. exceeding 50 tonnes ha^{-1} yr^{-1} or ca. 4 mm yr^{-1}) this means that effects of erosion on crop productivity will be relatively small on the decennial or centennial time scale, provided that nutrient losses due to erosion are compensated. Over longer time spans, however, the effect of these losses becomes very significant.

On the positive side, transported dust affects distant ecosystems, increasing plant productivity by providing nutrients not provided by the parent soil, as seen in Hawaii (Chadwick *et al.*, 1999) and the Amazon (Mahowald *et al.*, 2008). Transported dust can also provide chemical constituents that affect soil development, as seen in the development of terra rossa soils in Bermuda and Spain (Muhs *et al.*, 2010, 2012).

6.1.6 | Soil erosion and the environment

The direct negative effects of soil erosion are not limited to agriculture. The sediment produced by erosion also pollutes water streams with sediment and nutrients, thereby reducing water quality. In addition, sediment contributes to siltation in reservoirs and lakes. However, not all sediments trapped in reservoirs originate from agricultural land. Other processes such as bank erosion, landslides and natural surface erosion which contribute to reservoir sedimentation are also very important and are often dominant at large scales.

Wind erosion and dust transport have been studied for many years. For example, in 1646, Wendelin first described purple rain in Brussels that we now recognize as coloured dust transported to Europe from Africa (Wendelin, 1646 as cited in Stout, Warren and Gill, 2009). Charles Darwin studied dust that fell on the HMS Beagle in the 1830s and 1840s (Darwin, 1845, 1846) and the dust collected was found to contain viable microbes even today (Gorbushina *et al.*, 2007).

Wind erosion can originate from natural landscapes and from landscapes affected by anthropogenic (human-related) activities (Figure 6.4). Aeolian processes impact soil development, mineralogy, soil physical and biogeochemical properties, and redistribution of soil nutrients, organic materials, and sequestered contaminants. Wind erosion also affects landscape evolution, plant productivity, human and animal health (Ravi *et al.*, 2011), atmospheric properties including effects on solar radiation and cloud attributes (Shao *et al.*, 2011), air quality, and other factors (Field *et al.*, 2010; Ravi *et al.*, 2011). The effects of wind erosion occur at the field, landscape, regional, and global scales.

Figure 6.4 | Dust storm near Meadow, Texas, USA

At the field and landscape scales, wind erosion winnows the finer and more chemically active portion of the soil which carries biogeochemicals, including plant nutrients, soil carbon and microbial products. In some cases, wind erosion processes modify the surface properties by causing increases in sand content while reducing the soil water holding capacity and plant productivity (Zobeck and Van Pelt, 2011). Although some of this eroded sediment is deposited relatively close to field boundaries, often much of it enters into suspended mode and may be transported high in the atmosphere to travel great distances. This long-range transport of dust produces effects at the global and regional scales Atmospheric dust produced by wind erosion profoundly influences the energy balance of the Earth system by carrying organic material, iron, phosphorus and other nutrients to the oceans, affecting ocean productivity and subsequent ocean-atmosphere CO_2 exchange (Shao *et al.*, 2011).

6.1.7 | Effects of hydrology and water

Wind erodibility and subsequent erosion and dust emissions are affected by hydrology and water in several ways. Remote sensing studies of dust sources by Prospero *et al.* (2002) showed that many major dust sources originate from deep alluvial deposits formed by intermittent flooding during the Quaternary and Holocene. These sources, now in drylands, originated when water was more plentiful and produced an ample supply of wind-erodible sediment (Ginoux *et al.*, 2012). In many areas, particularly in areas with more limited erodible sediment supply, dust emissions increase after new inundations of ephemeral water supplies provide additional erodible sediment. However, many fluvial-related dust sources have also developed from the exposure, due to anthropogenic factors, of the bottoms of former lakes such as at Owens Lake in the United States (Reheis, 1997) and the Aral Sea Basin in Uzbekistan (Singer *et al.*, 2003). In these cases, usually water has been extracted from the lake for irrigation or human consumptive needs. This issue will be accentuated as increasing demand for water in dryland regions is met from reservoirs.

Near-surface soil water content has long been recognized to have a significant effect on the threshold wind velocity needed for wind erosion (Akiba, 1933; Chepil, 1956). Soil water acts to bind particles together to resist the shearing force of wind on the particles. In addition, soil water affects vegetative growth, which also affects wind erosion. Research has shown a time-dependent change in the controlling factors for sediment emission and transport from soil water to wind speed (Wiggs, Baird and Atherton, 2004). The change of controlling factors was found to be very sensitive to the prevailing water conditions and, for the sandy soil tested, took place in a very short period of time. They found the soil water content where wind erosion commenced was between 4 and 6 percent (Wiggs, Baird and Atherton, 2004). However, the effect of soil water on wind erosion of dry soils is also sensitive to changes in air relative humidity (Ravi *et al.*, 2006). Recent work on atmospheric dust concentrations have confirmed this sensitivity, finding that dust concentration increased with relative humidity, reaching a maximum around 25 percent and thereafter decreased with relative humidity (Csavina *et al.*, 2014). Climate-induced changes in hydrology and water may produce profound changes in wind erosion and dust emissions as the soil erodibility is altered.

6.1.8 | Vegetation effects

The effect of vegetation on wind erosion is complex. In native conditions, the wind influences patterns of vegetation and soils and these patterns, in turn, affect wind erosion at patch to landscape scales (Okin, Gillette and Herrick, 2006; Okin *et al.*, 2009; Munson, Belnap and Okin, 2011). In agricultural systems, the vegetation is manipulated by managers and its effects vary spatially and temporally from non-managed systems. The protective effects of vegetation are well known. A wide variety of methods and models has been devised to describe the protective effects of vegetation. In general, as vegetation height and cover increase, wind erosion of erodible land decreases. Vegetation affects wind erodibility by: (1) acting to extract momentum from the wind and thereby reducing the wind energy applied to the soil surface; (2) directly sheltering the soil surface from the wind by covering part of the surface and reducing the leeside wind velocity; and (3) trapping windborne particles, so reducing the horizontal and vertical flux of sediment (Okin, Gillette and Herrick, 2006). Trapping of sediment leads to redistribution of nutrients and modifies surface soil properties such as water infiltration rate and soil bulk density.

Vegetation cover affects nutrient removal, which in turn affects plant productivity. A study of the effects of grass cover on wind erosion in a desert ecosystem found increased wind erosion removed 25 percent of the total soil organic carbon and nitrogen from the top 5 cm of soil after only three windy seasons (Li *et al.*, 2007). Studies of agricultural crops on severely eroded cropland found 40 percent reductions in cotton and kenaf yields and 58 percent reduction in grain yield in sorghum (Zobeck and Bilbro, 2001). The eroded areas in this study had statistically significantly less phosphorus than the adjacent non-eroded areas. Climatic changes that reduce the cover of vegetation in drylands will increase wind erosion and dust emissions, and likely result in increased soil degradation and reduced plant productivity.

6.1.9 | Alteration of nutrient and dust cycling

Recognition of a dust cycle, along with other important cycles such as the energy, carbon and water cycles, has become an emerging core theme in Earth system science (Shao *et al.*, 2011). Dust cycles are dependent upon the soil and climate systems within which they operate. The dust cycle is a product, in part, of the soil system. As dust is transported globally, it interacts with other cycles by participating in a range of physical and biogeochemical processes. The dust carries important nutrients to otherwise sterile soils and so may improve productivity (Chadwick *et al.*, 1999; Mahowald *et al.*, 2008). Dust may also transport soil parent material (Reynolds *et al.*, 2006), trace metals (Van Pelt and Zobeck, 2007), soil biota (Gardner *et al.*, 2012) and toxic anthropogenics (Larney *et al.*, 1999) among ecosystems. Although the fact is not widely recognized, the global dust cycle is intimately tied to the global carbon cycle (Chappell *et al.*, 2013). Wind and water erosion both redistribute soil organic carbon within terrestrial, atmospheric and aquatic ecosystems. This carbon is selectively removed from the soil. This was recently demonstrated in an Australian study where the soil organic carbon in dust was from 1.7 to over seven times that of the source soil (Webb *et al.*, 2012). Changing climate will alter these cycles, producing complex and uncertain environmental effects.

6.1.10 | Trends in soil erosion

While rates of soil erosion are still very high on extensive areas of cropland and rangeland, erosion rates have been significantly reduced in several areas of the world in recent decades. The best documented example is the reduction of erosion rates on cropland in the United States. Average water erosion rates on cropland were reduced from 10.8 to 7.4 tonnes ha^{-1} yr^{-1} between 1982 and 2007, while wind erosion rates reduced from 8.9 to 6.2 tonnes ha^{-1} yr^{-1} over the same time span. Another example is the reduction of soil erosion in Brazil through the application of no-tillage from ca. 1980 onwards. This is estimated to have led to a reduction of erosion rates by 70-90 percent over large parts of Brazilian cropland. Studies have shown that erosion rates can be greatly reduced in nearly every situation through the application of appropriate management techniques and structural measures such as terrace and waterway construction (see, for example, Pansak *et al.*, 2008).

However, in many areas of the world, adoption of soil conservation measures is slow. While the reasons for this are diverse, a key point is that the adoption of soil conservation measures is generally not directly beneficial to farmers. This is as true in intensive mechanized systems in the West as it is for smallholder farming in the developing world. This is not surprising: the implementation of conservation measures does not, as such, directly increase yields or efficiencies while the detrimental effects of erosion on the soil capital only become visible over time scales that range from decades to centuries. Hence, farmers do not have a direct incentive to adopt soil conservation measures.

In some cases, this problem may be overcome through financial incentives or by regulation. It is clear, however, that this is not always possible. We need, therefore, to rethink our vision on soil conservation. Essentially, further adoption of soil conservation measures will not in the first place depend on refinement or optimization of technologies. Technology already exists to reduce erosion to acceptable levels under most circumstances. What is critically important is to work out *how to incorporate soil conservation in an agricultural system that, as a whole, increases the net returns of farmers*. In developing approaches that build in incentives to soil conservation, it is vital to account for local conditions, including the extent to which local markets can provide incentives to sustainable agriculture.

The potential for agricultural intensification is key here. In many areas around the world, crop yields are low and more land is cultivated than is strictly necessary. As a result, large tracts of steep, marginal land are at present used for agriculture without the implementation of proper soil conservation technology, with the result that these areas are subject to high erosion rates. Intensification of production on higher potential land is an option. This not only reduces extension into marginal, highly erodible areas but may also benefit biodiversity and overall carbon storage at the landscape scale.

Erosion can also be checked by forestation. In many areas there is now a net gain of forest area. This reforestation, which is largely of marginal land, is related to four main factors: agricultural intensification; diminishing need for firewood; an increase in exchange and trade making it possible to grow products in the most suitable areas; and an increased public awareness of the problems caused by deforestation. Development of conservation policies should consider these tendencies and stimulate them wherever possible.

6.1.11 | Conclusions

Soil erosion has been recognized as a main problem threatening the sustainability of agriculture for a long time and the magnitude of the problem can now be correctly quantified. The technology to reduce erosion now exists and, over the last decades, significant efforts have been to reduce erosion rates. These efforts have been partially successful. However, erosion rates are still high on much of the agricultural land of the globe, and this is related to the lack of economic incentives for today's farmers to conserve the soil resource for future generations. Tackling this problem requires the soil erosion problem to be reframed. Solutions need to be embedded in policies and programs that support the development of more sustainable agricultural systems.

6.2.1 | Introduction

An evaluation of the various functions of carbon (C) stored in the soil and its role in the global C-cycle require knowledge of the amount and geographic distribution of C stored in the soil. The functions of soil C are determined by the chemical and physical properties of the components that contain C. Chemical properties of soil C determine properties such as soil pH, nutrient storage and availability and regulating functions affecting the water cycle. Soil physical properties related to functions of C are: soil structure, particle agglomeration, and stability. These properties in turn influence water infiltration rates and resistance to water and wind erosion.

Soil C is separated into: (I) inorganic chemical substances (soil inorganic carbon: SIC), mainly carbonates; and (II) C as part of organic compounds (soil organic carbon: SOC). The amount of SIC in the first one meter of soil was estimated at 695 - 748 Pg carbonate C (Batjes, 1996). The C stored as SOC is about twice the C stored as SIC (1 502 Pg C; Jobbágy and Jackson, 2000). Carbonates are less susceptible to react to anthropogenic changes to the environment than are organic compounds. In addition, the amount and type of organic C compounds are interdependent with environmental conditions, such as land use and management practices. These two characteristics have led to the definition of the persistence of SOC as an ecosystem property (Schmidt *et al.*, 2011). Thus, assessments of soil C stocks and their spatial distribution often concentrate on SOC alone.

Although SOC mainly originates from plant material there is no simple correlation between the amount of C stored in the above-ground plant material and the SOC stocks (Amundson, 2001; Smith, 2012). In fact, the processes involved in decomposing organic material and their mineralization are complex and details are not yet fully understood (Schmidt *et al.*, 2011). However, the effects on SOC of anthropogenic activities of land use change and management practices are known. Given the large amount of C stored in soils and the possibility of influencing this amount through land management to act as a source or sink for atmospheric C, strategies for maintaining SOC have been devised. These strategies follow two main approaches: (I) seeking to enhance existing SOC by increasing the amount of biomass of the terrestrial biosphere; (II) seeking to decrease the loss of SOC by reducing the respiration rate (Smith *et al.*, 2008). To provide a quantitative appraisal of the possible gains or losses in SOC from measures taken either to increase the input of organic material or decrease losses of soil organic matter (SOM), an assessment of the current situation is needed.

Studies on historic developments in SOC stocks concentrate on the effect of changes in land use. These changes mainly concern the transformation of land from a natural state to an agricultural ecosystem, which in fact now covers more than one third of the global terrestrial area. For the conversion of forest to cropland, losses in SOC stocks of 25-30 percent were observed for temperate regions, with higher losses recorded for the tropics. Estimates of future trends in SOC stocks concentrate on the effect of changing climatic conditions on rates of organic matter accumulation and decomposition. As options for changes in land use are relatively limited, approaches to mitigation of climate change effects have focussed largely on management practices.

6.2.2 | Estimates of global soil organic carbon stocks

It is important to know past, current and likely future SOC stocks because of their importance to climate change and food security. When assessing the amount of C stored in the soil, studies often concentrate on C contained in dead and decomposed organic material or in organic matter located within the soil profile to a given depth and for a specific area. The mass of C stored in the SOM is also termed SOC stocks. SOC stocks are computed as a function of organic C-content, bulk density, depth and the amount of soil remaining after removing the volume taken up by coarse fragments in a unit of volume. Any of these factors can introduce uncertainties to the estimates.

Global estimates of SOC stocks have been published for many decades. One of the earliest estimates was published in 1951 (Rubey, 1951), indicating a global SOC stock of 710 Pg C. This estimate remained current for 25 years until the FAO-UNESCO soil map became available. Analysis of the map data led to a much higher estimate of 3 000 Pg C in the soil (Bohn, 1976). Several studies of global SOC stock followed with varying estimates (Bohn, 1982: 2 200 Pg organic C; Post *et al.*, 1982: 1 395 Pg organic C) An estimate of 1 576 Pg of SOC to 1 m depth was put forward by Eswaran, Van Den Berg and Reich, 1993. Global SOC stocks to 1 m depth of 1 462 – 1 548 Pg of SOC were estimated by Batjes, 1996. An estimate of 1 502 Pg organic C for the first 1 m of soil is often used (Jobbágy and Jackson, 2000; Batjes, 2002). The estimate of 1 500 Pg of SOC for the top 1 m of soil was adopted by the IPCC (IPCC, 2000). Current global estimates, derived from the *Harmonized World Soil Database* (HWSD; FAO/IIASA/ISRIC/ISS-CAS/JRC, 2009), suggest that approximately 1 417 Pg of SOC are stored in the first meter of soil and about 716 Pg organic C in the top 30 cm (Hiederer and Köchy, 2011).

Fewer estimates of global SOC stock estimates are available for a depth below 1 m. Global SOC stocks to a depth of 3 m are estimated at 2 344 Pg of SOC (Jobbágy and Jackson, 2000) or 3 000 Pg of SOC (Jansson *et al.*, 2010). Recent estimates of SOC in Cryosols may further increase the estimates of global SOC stocks (Tarnocai *et al.*, 2009).

In a comparison of 27 studies on global SOC stock published between 1951 and 2011, the estimates published were found to range from 504 to 3 000 Pg of SOC (Scharlemann *et al.*, 2014). The median of all published estimates is 1 460 Pg of SOC to 1 m depth. Spatially explicit estimates were found to span over 1 965 Pg of SOC. Large uncertainties over SOC stocks concern Histosols since soil data are often limited to a depth of 1 m (Eswaran, Van Den Berg and Reich, 1993). Particularly affected are the soils of the Arctic (Tarnocai *et al.*, 2009) and peatlands in South Asia (Couwenberg, Dommain and Joosten, 2010). The range in the estimates of global SOC stocks correspond to or exceed the amount of C held in the atmosphere, which was estimated at 720 Pg C (Falkowski *et al.*, 2000) and at 820 Pg C for present conditions (Mackey *et al.*, 2013).

With respect to the uncertainty in the estimates of global SOC stocks, various approximations are observed. For an estimated SOC stock of 1 395 Pg of SOC Post *et al.* (1982) assume a standard deviation of ± 200 Pg organic C, provided that the SOC density data are the only source of uncertainty. For the estimate of 1 502 Pg organic C to 1 m depth, Jobbágy and Jackson (2000) suggested an error of the mean of ± 320 Pg C at 1 standard deviation, provided that the SOC content data are the only source of uncertainty. The different assumptions on the causes of uncertainty between the studies (SOC density or content) are quite significant. Based on the HWSD, Todd-Brown *et al.* (2013) provide an interval of estimated global SOC stock of 890 to 1 660 Pg of SOC to a depth of 1 m with a 95 percent confidence level. This range corresponds to approximately 385 Pg SOC at 2 standard deviations from the mean.

With a small number of large-scale data sets available, the variations in SOC stock estimates may be attributed to the analysis method applied as much as to the data used. It also implies that various global SOC stock estimates are not independent and that the variability in the estimates could not necessarily be reduced by an increase in the number in such estimates.

One problem common to all large databases is that the properties were assessed decades ago and stretched over long periods. For example, the DSMW or the ESDB, of which components are included in the HWSD, originated from data collected during the 1950s and 1960s. With the dependence of SOC on climatic conditions and anthropogenic activities, SOC stocks established decades apart are likely to represent significantly different levels, notably in areas where changes in land use or management occurred, such as conversion of natural grassland and forest to agricultural land or urban areas. In extreme cases draining peatlands can lead to a loss of organic material to the degree that the soil no longer qualifies as peat because the organic C content decreases below 12 percent content and the thickness of the remaining organic layer is less than 40 cm (FAO/ISRIC/ISSS, 1998). An example of this change is given by the agricultural areas in north-eastern Netherlands, where the drainage in the 1960's of areas previously classified as peatland caused the SOC content to fall to 7.5 percent (Panagos *et al.*, 2013). Without further adjustments of SOC stock estimates to take account of local changes in the factors that influence SOC, no clear timestamp can be attached to the global estimates. This lack of a clear timestamp of SOC stocks is of consequence when estimating temporal changes in SOC stocks. Estimates of changes in SOC stock therefore concentrate on modelling variations in SOC from changes in land use and cover.

6.2.3 | Spatial distribution of SOC

Different methods of combining point data from soil profiles with soil spatial layers and ancillary ecological data can be applied to derive spatial estimates of SOC stocks (Kern, 1994). SOC density and stock estimates from soil profile data were combined with spatial data of major ecosystems by Post *et al.* (1982). The total SOC stocks for all life zones to a depth of 1 m was 1 395 Pg of SOC. A combination of soil profile data with ancillary information on climate, vegetation and land use was used by Jobbágy and Jackson (2000) to estimate SOC stocks in 11 biomes. The estimates for the biomes were further divided into increments of 1 m soil depth and of 20 cm for the first meter. The distribution of SOC stocks by ecological regions has also been presented, for example by Amundson (2001), who used life zones as the study unit. Eglin *et al.* (2010) used the SOC stock estimates to a depth of 3 m from Jobbágy and Jackson (2000) and modified SOC stocks in permafrost areas (Tarnocai *et al.*, 2009). These SOC stock estimates were combined with estimates provided by the IPCC (IPCC, 2000) of C in vegetation to derive estimates of C in soil and biomass for 10 biomes, with an explicit class for peatlands. A step towards adding a temporal dimension to spatial SOC stock estimates, assessing historical and future trends, was made possible by the availability of SOC models. Combining the models with historic land use and climate data has allowed estimation of SOC stocks with a timestamp and with regional variations (Eglin *et al.*, 2010; Schmidt *et al.*, 2011).

Carré *et al.* (2010) produced estimates of SOC stocks and density using climate data, IPCC methodology and the Harmonized World Soil Database. The results by IPCC Climate Region are presented in Table 6.1.

Table 6.1 | Distribution of Soil Organic Carbon Stocks and Density by IPCC Climate Region

** Differences in topsoil and subsoil sum are due to data rounding*
*** Total includes 1.4 Pg C in undefined climate regions*

IPCC Climate Region	IPCC	HWSDa			
	Tier 1	Topsoil	Subsoil	Soil	Density
	0-30 cm	0-≤30 cm	30-≤100 cm	0-≤100 cm	0-≤30 cm
	Pg C	Pg C	Pg C	Pg C*	Mg C ha^{-1}
Tropical Wet	52.4	62.6	65.4	128.0	66.5
Tropical Moist	94.5	78.6	72.3	150.9	45.0
Tropical Dry	99.9	67.3	69.0	136.2	22.0
Tropical Montane	49.8	29.6	26.5	56.1	40.3
Warm Temperate Moist	41.7	33.3	29.7	63.0	60.2
Warm Temperate Dry	42.9	38.9	39.6	78.5	30.8
Cool Temperate Moist	110.6	104.1	106.2	210.3	88.2
Cool Temperate Dry	56.9	52.2	50.0	102.2	42.7
Boreal Moist	137.3	162.0	194.7	356.7	117.6
Boreal Dry	30.3	32.0	37.0	68.1	84.0
Polar Moist	26.8	30.6	21.7	52.4	40.4
Polar Dry	7.2	8.0	4.3	12.3	40.5
Total	**750.3****	**699.3**	**716.4**	**1415.7**	**52.1**

The table shows that according to the processed data from HWSD, most SOC (356.7 Pg C) is stored in the '*Boreal Moist*' climatic region. The second largest stock is found in the '*Cool Temperate Moist*' region (210.3 Pg C). With 117.6 Mg C ha^{-1} and 88.2 Mg C ha^{-1}, these climate regions also have the highest SOC densities. These figures compare poorly with those presented by Post *et al.* (1982). A major source for the deviation is the difference in the definition of the life zones as compared to the climatic regions, which lead to the delineation of different areas.

Using the IPCC Tier 1 default values for organic C in mineral soils and retaining the stocks for organic soil gives global organic C stock in the upper 30 cm of soil of 750.3 Pg C. This estimate is 51 Pg C (7.3 percent) higher than the estimates derived from the HWSDa topsoil layer.

When comparing the two spatial SOC stock estimates by IPCC climate region, the stocks within each region are broadly similar. A notable difference is for soils in the '*Tropical Dry*' climate region. The IPCC Tier 1 SOC map gives 99.9 Pg C for this zone, compared to 67.3 Pg C found in the HWSDa.

In the interpretation of the figure for SOC stocks of the IPCC Tier 1 Soil Organic Carbon layer, it should be considered that organic soils were only added to the mineral soil layer in places where this soil type is not found in association with mineral soils. Using all organic soil data is likely to increase the global SOC stocks. However, the Tier 1 default values are calculated over a constant depth of 30cm, although some soils are shallower, which in turn would reduce the stocks.

6.2.4 | Spatial distribution of carbon in biomass

A global map of C stored in biomass following the IPCC Tier 1 approach was produced by the European Commission Joint Research Centre (Carré at al., 2010; EU, 2004; Hiederer *et al.*, 2010). The C stocks are determined for above- and below-ground biomass and include dead organic matter for the relevant vegetation types. The default factors largely follow the IPCC specification, with specific attention given to agricultural areas. The underlying vegetation data are based on the GlobCover V2.2 (ESA, 2011). Because the GlobCover data limits cropland to areas below 57º N in Europe the data were merged with the M 3-Cropland (Ramankutty *et al.*, 2008) and Crops (Monfreda, Ramankutty and Foley, 2008). In a comparison of the geographic distribution of IPCC vegetation classes between the GlobCover and the M 3 Cropland and Pasture data, some notable differences were identified (Hiederer *et al.*, 2010). Some of the differences were attributed to the dissimilar definition of the vegetation classes in the data sets, although others, such as the separation of shrub land from open forest or confusion between cropland and pastures, seem to be the result of the classification algorithm used or of sensor characteristics.

The global biomass map thus generated by the Joint Research Centre (JRC) estimates the storage of C in the above-ground and below-ground vegetation and dead organic matter to be 456 Pg C. The JRC estimates are thus 44 Pg C (8.8 percent) lower than those of the '*New IPCC Tier-1 Global Biomass Carbon Map for the Year 2000*' (Ruesch and Gibbs, 2008). The difference is not evenly distributed between geographic regions. A comparison of carbon in C by climatic region is given in Figure 6.5.

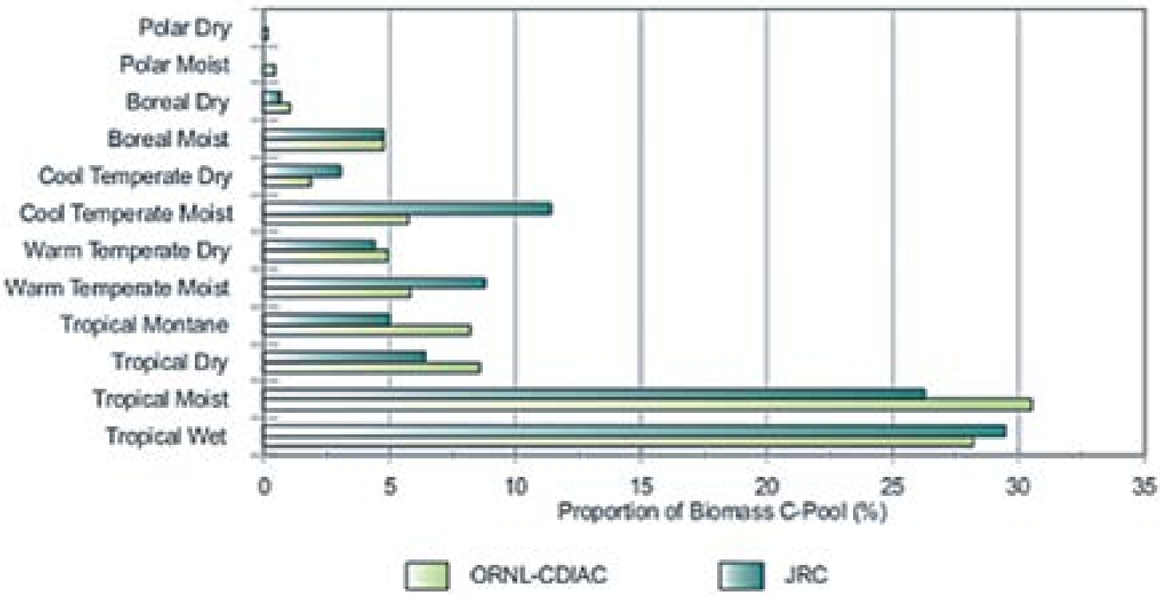

Figure 6.5 | Distribution of carbon in biomass between ORNL-CDIAC Biomass and JRC Carbon Biomass Map

The graph shows that the ORNL-CDIAC Biomass and the JRC Carbon Biomass map are mostly comparable, but the JRC map places relatively more C in the biomass in '*Cool Temperate Moist*' (11.4 percent of the total C stock in biomass; 51.8 Pg C) and '*Warm Temperate Moist*' (8.7 percent of the total C stock in biomass; 39.9 Pg C) climate regions at the expense of other regions. By contrast, the ORNL-CDIAC Biomass map locates only 5.7 percent of the total C stock in biomass (28.4 Pg C) in the '*Cool Temperate Moist*' and 5.7 percent of the total C stock in biomass (28.7 Pg C) in the '*Warm Temperate Moist*' climate region.

For the total terrestrial pool of organic C, biomass is the more important pool only in the climate regions '*Tropical Wet*' and '*Tropical Moist*'. For all other climatic regions, the soil stores more organic C than the biomass (Scharleman *et al.*, 2014).

6.2.5 | Distribution of terrestrial carbon pool by vegetation class

Areas where SOC or biomass C dominate could be identified by computing the difference between the two layers. The resulting layer is presented in Figure 6.6.

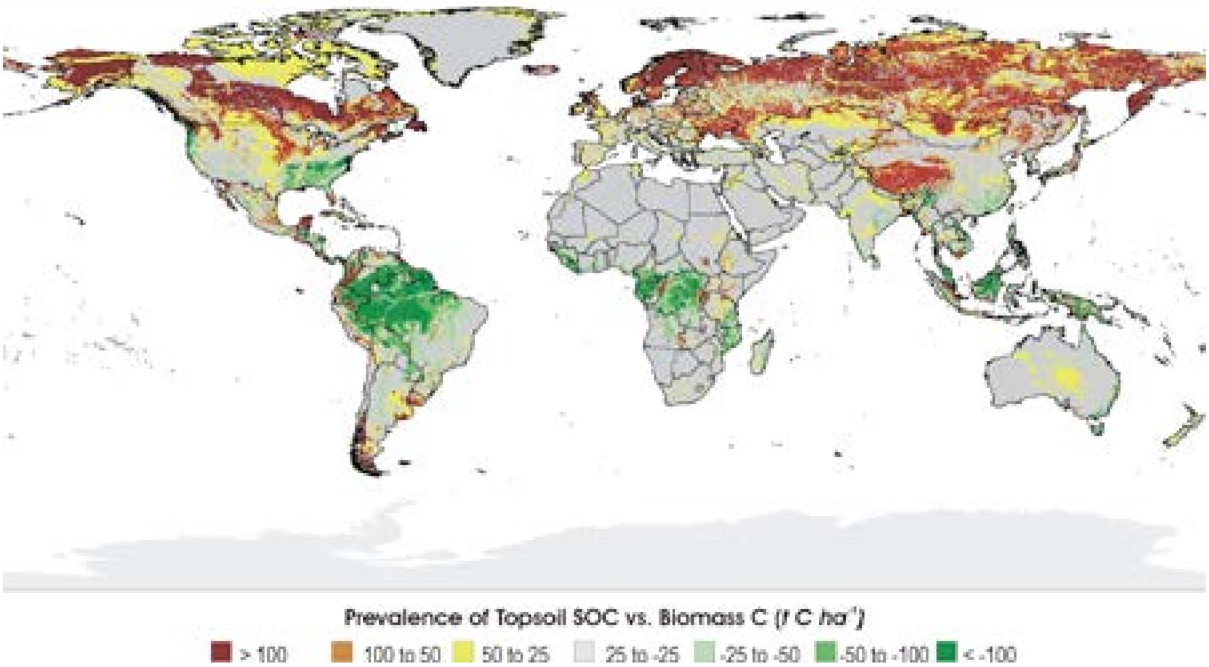

Figure 6.6 | Prevalence of carbon in the topsoil or biomass

The figure shows that, as a general propensity, soil dominates the terrestrial C pool in cooler climates while vegetation forms the dominant pool of terrestrial C in tropical regions.

In an attempt to provide C stock estimates for broad land use activities, global GLC 2000 data layers were used. The GLC 2000 categories were re-classified according to the assignments for these classes given by Ruesch and Gibbs (2008). A difference in the assignment was applied to GLC 2000 classes 16 (*Cultivated and Managed Areas*) and 23 (*Irrigated Agriculture*). In the broad classification these areas were grouped with other areas mainly associated with an absence of soil or biomass (bare areas, glaciers, etc.). For the analysis of the distribution of organic C, a separate class of '*Agriculture and Managed Areas*' was created by merging the GLC2000 classes 16 and 23. For each of the broad vegetation classes the organic C stock was extracted by pool. The results are presented in Table 6.2.

Table 6.2 | Distribution of terrestrial organic carbon by stock and broad vegetation class

Vegetation Classes	Topsoil	Subsoil	Soil		Biomass		Terrestrial C Stock	
	Pg C	Pg C	Pg C	percent	Pg C	percent	Pg C	percent
Broadleaf Forest	124.7	112.4	237.1	16.8	272.2	54.4	509.4	26.6
Evergreen Forest	126.8	139.7	266.4	18.8	46.4	9.3	312.9	16.3
Mixed Forest	40.5	47.8	88.3	6.2	21.8	4.4	110.1	5.7
Burnt Forest and Natural Forest Mosaic	27.4	36.2	63.6	4.5	10.9	2.2	74.5	3.9
Forest/Cropland Mosaic	23.2	23.4	46.6	3.3	28.0	5.6	74.6	3.9
Forest	342.6	359.5	702.0	49.6	379.3	75.9	1081.5	56.4
Shrub Cover	89.2	102.4	191.6	13.5	51.8	10.4	243.4	12.7
Grasslands	60.5	52.1	112.6	8.0	18.0	3.6	130.5	6.8
Sparse Grassland and Grassland Mosaic	69.0	65.5	134.5	9.5	12.7	2.5	147.2	7.7
Grassland	218.7	220.0	438.7	31.0	82.5	16.5	521.1	27.2
Agriculture and managed areas	80.8	79.4	160.2	11.3	26.7	5.3	186.9	9.8
Other Classes	57.3	57.4	114.7	8.1	11.4	2.3	126.2	6.6

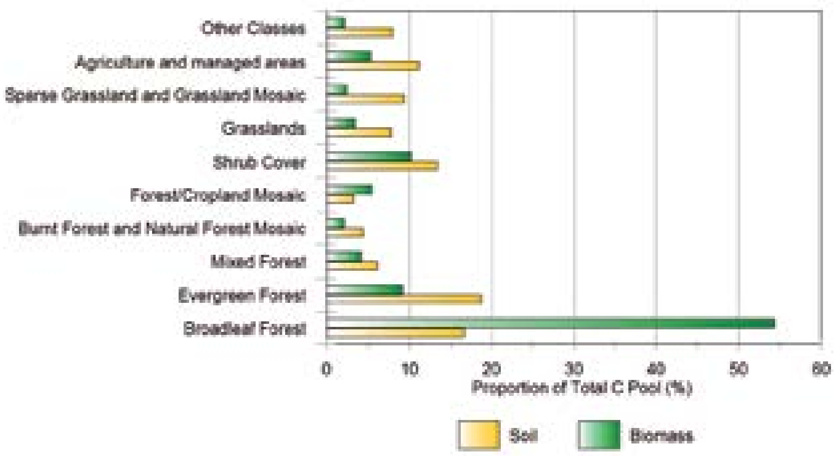

Figure 6.7 | Proportion of carbon in broad vegetation classes for soil and biomass carbon pool

The single largest stock for terrestrial C is attributed to areas with broadleaf forest (509.4 Pg C). This forest type contains approximately one quarter of all terrestrial organic C in either the soil or the biomass.

The proportions of the C stored in the soil and biomass stocks by broad vegetation class is graphically presented in Figure 6.7.

For the biomass C stock alone, broad forests account for over 50 percent of the C in that pool, but only 16.8 percent of the organic C is stored in the soils under this vegetation type. With the exception of the 'Forest/ *Cropland Mosaic*', in all other vegetation classes the soil stores more C than the biomass.

6.2.6 | Historic trends in soil carbon stocks

The SOC stocks are more susceptible to anthropogenic activities and natural factors than are SIC stocks. Conversion of natural to agro-ecosystems in the past has led to decline in the SOC stock of the surface layers and also in SOC in the total profile for most soils. The magnitude of the historic loss, however, differs among soils and climates. The magnitude and rate of loss are higher for soils within the tropics than for those of temperate climates. Losses are also higher for coarse-textured than for heavy-textured soils, higher for soils containing higher SOC stocks, and higher for soils under subsistence or 'extractive' farming than for those farmed with more science-based agricultural practices. Depletion is also exacerbated by drainage of wetlands, by ploughing, and by biomass burning or removal. Some soils in the tropics can lose 50 percent of their previous pool within five years following deforestation and conversion to agricultural land use. The rate and magnitude of SOC loss are exacerbated in soils vulnerable to accelerated erosion, salinization, nutrient depletion or imbalance, structural decline and compaction, acidification, elemental toxicity, pollution and contamination.

Estimates of the magnitude of historic SOC loss vary widely. The historic loss has been estimated at 40 Pg by Houghton (1995), 55 Pg by IPCC (1995) and Schimel (1995), 150 Pg by Bohr (1978), 500 Pg by Wallace (1994) and 537 Pg by Buringh (1984). The average of these estimates is 223 Pg C year^{-1}. Lal (1999) estimated the magnitude of SOC loss since 1850 at 47 to 104 Pg for different biomes (Table 6.3); 66 to 90 Pg for major soils (Table 6.4); and 19 to 31 Pg by erosional processes (Table 6.5). While the historic loss from *Gelisols* or permafrost soils is zero, these soils, which contain a vast amount of SOC reserves, are vulnerable to projected warming and the attendant positive feedback.

Table 6.3 | Estimate of the historic SOC depletion from principal biomes. Source: Lal, 1999.

Biome	Change in Area 10^6 ha	Historic SOC Loss Pg C
Forests	1300	23 - 53
Woodlands	180	3 - 7
Shrublands	140	1 - 4
Grasslands	660	20 - 40
Total		**47 - 104**

Table 6.4 | Estimates of historic SOC depletion from major soil orders. Source: Lal, 1999; Hillel and Rosenzweig, 2009.

Soil Order	Historic Area 10^6 ha	Present SOC Pool Pg C	Historic SOC Loss Pg C
Alfisols	1330	91	15 - 18
Andisols	110	30	5 - 7
Aridsols	1560	54	0.2 - 0.3
Entisols	2170	232	0.8 - 1.3
Histosols	160	312	?
Inceptisols	950	324	8 - 13
Mollisols	920	120	7 - 11
Oxisols	1010	99	22 - 27
Sposòsols	350	67	1 - 3
Ultisols	1170	98	6 - 7
Vertisols	320	18	1 - 2
Gelisols	1120	238	0
Others	1870	17	0.2 - 3
Total	**13050**	**1700**	**66 - 90**

Table 6.5 | Estimates of historic SOC loss from accelerated erosion by water and wind. Source: Lal, 1999.

Erosion	Area		Historic SOC Loss Pg
	Water 10^6 ha	Wind 10^6 ha	
Light	343	269	2 - 3
Moderate	527	254	10 - 16
Strong	224	26	7 - 12
Total			**19 - 31**

Estimates of the historic C loss are useful as a reference point for assessing the technical potential of C re-sequestration in soil. While the loss of SOC can be rapid, especially in soils of the tropical ecosystems, the rate of re-carbonization is extremely slow. The slow rate of re-sequestration is a major challenge to identifying appropriate land use and to promoting adoption of soil/water/animal/plant management systems that could create a positive soil/ecosystem C budget.

6.2.7 | Future loss of SOC under climate change

Projected changes in climate (temperature and precipitation) are likely to affect the SOC stock both directly and indirectly. Directly, the rate of decomposition by microbial processes is affected by both soil temperature and moisture regimes. Indirectly, changes in climate affect plant growth, net primary productivity, above and below-ground biomass, and the type and amount of residues with differential amounts of materials recalcitrance. Further, the rate and susceptibility to accelerated erosion, salinization and other degradation processes may be exacerbated by an increase in frequency of extreme events. Indeed, climate change can impact several soil forming factors, including rainfall, temperature, micro-organisms/biota and vegetation, thus affecting the rate of SOC accumulation (Jenny, 1930). Climate change may also alter species composition, and the rate of litter fall. However, disagreement exists regarding the effect of warming on SOC stock.

The annual rate of litter return, on which the rate of SOC accretion depends, varies among biomes (White, 1987; Grunwald, 1999). The rate of litterfall (Mg ha^{-1} yr^{-1}) is estimated at 0.1 to 0.4 for alpine and arctic regions, 2 - 4 for temperate grassland, 1.5 - 3 for coniferous forest, 1.5 - 4 for deciduous forest, 5 - 10 for tropical rainforest, and 1 to 2 for arable land (White, 1987). Increase in soil temperature may exponentially increase the rate of soil respiration (Tóth *et al.*, 2007; Lenton and Huntingford, 2003). However, because of increase in the number and activity of soil fungi in the warmer soil, there may also be increase in the relative amount of lignin and other recalcitrant compounds (Simpson *et al.*, 2007). The SOM decomposition is also more temperature-sensitive at low than at high temperature (Kirschbaum, 1995, 2000, 2006).

Thus, knowledge about the temperature–sensitivity of diverse SOC fractions, and their change in the soil under climate change, is important. Change in temperature by 1º Celsius may decrease the turnover times of 4-11 percent and 8-16 percent for the intermediate and stabilized fractions, respectively (Hakkenberg *et al.*, 2008). The decomposition rate is also influenced by the presence of physicochemical protection mechanisms (Conant *et al.*, 2011), especially occlusion within aggregates and by association with mineral surfaces (Freedman, 2014). It is argued that CO_2 emissions from soil response to climate warming are over-estimated, because the decomposition of old SOM is tolerant to temperature (Liski *et al.*, 1999). Thus, the effects of warming on SOM decomposition are governed by complex and interactive factors, and are difficult to predict. Despite much research, no consensus has yet emerged on the temperature sensitivity of SOM decomposition (Davidson and Janssens, 2006).

6.2.8 | Conclusions

Global SOC stocks have been estimated at about 1 500 Pg C for the topmost 1 m. However, a large uncertainty attaches to this estimate, which cannot easily be assigned to a specific period in time. Local variations may also be high, for example for SOC stocks in arctic regions and peatlands. Estimates of SOC stocks below 1 m depth are still evolving, with a tendency for more recent estimates to be higher than previous values. Estimates of the historic loss of SOC pools are also highly variable, ranging from 40 to 537 Pg. The global loss of SOC pool since 1850 is estimated at about 66±12 Pg. The projected response of SOC stock to climate change is a debatable issue. While an increase in temperature may increase the rate of respiration at low soil temperature, it may also shift microbial populations to fungi, increase relative proportions of lignin and other recalcitrant fractions, and increase protective mechanisms such as aggregation and reaction with mineral surfaces.

6.3.1 | Introduction

Soil contamination as a result of anthropogenic activities has been a widespread problem globally (Bundschuh *et al.*, 2012; DEA, 2001; EEA, 2014; Luo *et al.*, 2009; SSR, 2010). Soil contamination can be local or diffuse. Local soil contamination occurs where intensive industrial activities, inadequate waste disposal, mining, military activities or accidents introduce excessive amounts of contaminants. Diffuse soil contamination is the presence of a substance or agent in the soil as a result of human activity and emitted from dispersed sources. Diffuse contamination occurs where emission, transformation and dilution of contaminants in other media have occurred prior to their transfer to soil. The three major pathways responsible for the introduction of diffuse contaminants into soil are atmospheric deposition, agriculture, and flood events. These pathways can also cause local contamination in some instances. Causes of diffuse contamination tend to be dominated by excessive nutrient and pesticide applications, heavy metals, persistent organic pollutants and other inorganic contaminants. As a result, the relationship between the contaminant source and the level and spatial extent of soil contamination is indistinct.

While some soil degradation processes are directly observable in the field (erosion, landslides, sealing or even decline of organic matter), soil contamination as well as soil compaction or decline in soil biodiversity cannot be directly assessed, which makes them an insidious hazard. Moreover, diffuse contamination is linked to many uncertainties. The diversity of contaminants (particularly of the persistent organic pollutants, which are in constant evolution due to agrochemical developments) and the transformation of organic compounds in soils by biological activity into diverse metabolites make soil surveys to identify contaminants difficult and expensive. The effects of soil contamination also depend on soil properties, as these have an impact on the mobility, bioavailability, residence time and levels of contaminants. Direct effects of pollutants may not be immediately revealed because of the capacity of soils to store, immobilize and degrade them. Effects can, however, suddenly emerge after changes such as changes in land use that may alter environmental conditions (Stigliani *et al.*, 1991 - see also Chapter 7 on processes impacting service provision). Contaminants include inorganic compounds such as metallic trace-elements and radionuclides, and organic compounds like xenobiotic molecules. The application of some organic wastes to soils – for example, untreated biosolids – also increases the risk of spread of infectious diseases. A new challenge is that the so-called 'chemicals of emerging concern' (CECs) – for example, veterinary and human therapeutic agents such as antibiotics and hormones – are present in amendments added to soils, such as manures. These CECs can have an adverse effect on ecosystems and on human health (Jjemba, 2002; Osman, Rice and Codling, 2008; Jones and Graves, 2010).

6.3.2 | Global status of soil contamination

In most developed countries, waste disposal and treatment, industrial and commercial activities, storage, transport spills on land, military operations, and nuclear operations are the key sources of local soil contamination. Management of local soil contamination requires surveys to seek out sites that are likely to be contaminated, site investigations where the actual extent of contamination and its environmental impacts are defined, and implementation of remedial and after-care measures. By contrast, diffuse soil contamination is much harder to manage: in many instances it is not directly apparent but it may cover very large areas and represent a substantial threat. Despite the fact that most developed countries have implemented long-term soil surveys, even these countries still lack a harmonized soil monitoring system, and the real extent of diffuse soil contamination is not known.

According to the most recent data provided by the European Environmental Agency (EEA, 2014), total potentially contaminated sites in Europe are estimated to be more than 2.5 million, of which 340 000 are thought to be actually contaminated. Approximately one third of the high risk sites have been positively identified as contaminated, and of these only 15 percent have so far been successfully remediated (EEA, 2014). While trends vary across Europe, it is clear that the remediation of contaminated sites is still a significant undertaking. Waste disposal and industrial activities are the most important sources of soil contamination overall in Europe. The most frequent contaminants are heavy metals and mineral oils (EEA, 2014).

In the United States, sites contaminated with complex hazardous substances that impact soil, groundwater or surface water are placed on the Superfund National Priorities List (NPL). As of September 29, 2014, there were 1 322 final sites on the NPL. On 1 163 of these sites, measures to address the contamination threat have been completed. An additional 49 sites have been proposed. In addition, the Office of Solid Waste and Emergency Response (OSWER) has cleaned up over 540 000 sites and 9.3 million ha of contaminated land, all of which can be put back into use. In Canada, a total of 12 723 soil contaminated sites has been identified, with 1 699 sites related to surface soil contamination (Treasury Board of Canada Secretariat, 2014). The key soil contaminants include metals, petroleum hydrocarbons (PHCs), and polycyclic aromatic hydrocarbons (PAHs).

The pattern of contamination in Australia is similar to that of other developed countries. Industry, including the petroleum industry, mineral mining, chemical manufacture and processing facilities, and agricultural activities with their use of P fertilizer and pesticides, have caused soil contamination with heavy metals, hydrocarbons, mineral salts, particulates, etc. The total number of contaminated sites is estimated at 80 000 across Australia (DECA, 2010), with approximately 1 000 actual or potentially contaminated sites in South Australia (SKM, 2013).

Developing countries are undergoing significant industrialization. If appropriate legal and regulatory frameworks and enforcement capability are not developed, this may lead to soil contamination and pose risks to the environment and human health. In large conurbations, there is also a need for adequate provision of sanitation and drainage so that household wastes are collected and managed safely.

Asian countries experience considerable contamination of agricultural soil and crops by trace elements, and this contamination is becoming a threat to human health and the long-term sustainability of food production in the contaminated areas. In China, it is estimated that nearly 20 million ha of farmland (approximately one fifth of China's total farmland) is contaminated by heavy metals (Wei and Chen, 2001). This may result in a reduction of more than 10 million tons of food supplies each year in China (Wei and Chen, 2001). Atmospheric deposition (mainly from mining, smelting and fly ash) and livestock manures are the main sources of trace elements contaminating arable soil (Luo *et al*., 2009). Among the different trace elements contaminating Chinese agricultural soils, Cd is the biggest concern. Due to its high mobility in the soil (except in poorly drained soil where sulphides are present), it can be easily transferred to the food chain and so poses risks to human health. Arsenic is also naturally present in groundwater in many regions of Southeast Asia. This represents a threat to agriculture, particularly in rice paddy fields where anaerobic conditions prevail (Smedley, 2003; Hugh and Ravenscroft, 2009). Asia is also the largest contributor to the atmosphere of anthropogenic Hg, which originates from the chemical industry, from Hg mining and from gold mining (Li *et al*., 2009). All across Asia, areas under rapid economic development are experiencing moderate to severe contamination by heavy metals (Ng, 2010).

In many parts of Latin America, the results of anthropogenic activities, such as tailings and smelting operations in mining areas, have resulted in arsenic contamination in the soil. These operations enhance the mobilization of arsenic and cause adverse environmental impacts (see Section 4.3). Also in Latin America, the problem of arsenic contamination in water has been identified in 14 of the continent's 20 countries: Argentina, Bolivia, Brazil, Chile, Colombia, Cuba, Ecuador, El Salvador, Guatemala, Honduras, Mexico, Nicaragua, Peru and Uruguay. The number of exposed people in these countries is estimated to be about 14 million (Bundschuh *et al.*, 2012; Castro de Esparza, 2006). It is also estimated that during the late 1980s and early 1990s, 3 000 to 4 000 tonnes of Hg were deposited in the Amazon basin as a result of artisanal gold-mining activities, mainly in Brazil, Bolivia, Venezuela and Ecuador (de Lacerda, 2003). In addition, intensive use of fertilizers and pesticides in many parts of Latin America contributes to soil contamination and causes a range of environmental pollution and human health problems (UNEP, 2010).

In Africa, soil contamination has resulted from mining, spills, and improper handling of waste (Gzik *et al.*, 2003; SSR, 2010; EA, 2010). The Nigerian federal government reported more than 7 000 spills between 1970 and 2000. In Botswana and Mali, over 10 000 tonnes of pesticides, including DDT, aldrin, dieldrin, chlordane and heptachlor, have leaked from damaged containers and contaminated the soil (SSR, 2010). Soil contamination in the Near East and North Africa is linked to oil production and heavy mining. In arable land, a common source of soil pollution is the use of contaminated groundwater or wastewater for irrigation.

6.3.3 | Trends and legislation

In developed countries, legislation on contaminated land and the related regulatory mechanisms are well established. As a result, the extent of contaminated land is thoroughly reported. The European countries have created a common framework in the Thematic Strategy on Soil Protection (COM (2006) 231), which aims at sustainable use of soil, preservation of soil as a resource, and remediation of contaminated soil. The EC has also created networks such as CLARINET, NICOLE and SNOWMAN (Vicent, 2013). Investigations of suspected contaminated sites continue in Europe and as a result the total of contaminated sites listed is expected to increase by 50 percent by 2025 (EEA, 2007, 2012; EC, 2013). The number of remediated sites is expected to grow as well. In addition, regulation now requires industrial plants to control their wastes and prevent accidents, limiting the introduction of contaminants into the environment. As noted above, the United States has introduced a regulatory regime and has made significant progress on site clean-up.

In Asia, early legislation on contaminated land management (CLM) focused on contamination of agricultural land caused by industrialization and urbanization. Thus Japan, Taiwan, Province of China and South Korea have developed comprehensive CLM frameworks of laws, regulations and guidelines. Other Asian countries, however, are still at early stages of developing a CLM framework (Ng, 2010).

Atmospheric deposition (Section 4.4.1) is an important input of pollutants (Lofts *et al.*, 2007) and air quality regulation to decrease the load of contaminants on soils is therefore important. In most developed countries, relevant legislation is well established. In the case of long-range atmospheric pollution, international agreements are needed. In this regard, the Convention on Long-Range Transboundary Air Pollution (LRTAP) was signed in 1979. Conceived in response to the detrimental impact of acid rain in Europe, the Convention entered into force in 1983. Over the past 30 years, the Convention has been extended by eight further protocols that target pollutants such as S, NO_x, persistent organic pollutants, volatile organic compounds, ammonia and heavy metals. More recently, a global treaty to protect human health and the environment from the adverse effects of mercury - the 2013 Minamata Convention on Mercury - has been established.

CECs require due attention and they can include, but are not limited to, nanoparticles, pharmaceuticals, personal care products, estrogen-like compounds, flame retardants, detergents, and some industrial chemicals (including those in products and packaging) with potential significant impact on human health and aquatic life (Jones and Graves, 2010). Electronic waste (also referred to as 'e-waste') is of great concern given the increasing volumes generated each year, the hazardous nature of some of the components, and the exportation of this waste from industrialized countries to recycling centres in China, India and Pakistan (UNEP DEWA/GRID-Europe, 2005). This chain risks violating the Basel Convention on the Control of Transboundary Movements of Hazardous Wastes and their Disposal, which was adopted in 1989 and came into force in 1992.

Recently some countries have implemented policies and programmes to encourage waste minimization. These programmes of 'Extended Producer Responsibility' make producers responsible for the costs of managing their products at the end of their life. This approach is expected to encourage the manufacture of more environmentally-friendly electronic products (UNEP DEWA/GRID-Europe, 2005).

6.4 | Soil acidification status and trends

6.4.1 | Processes and causes of acidification

Soil acidity increases with the build-up of hydrogen (H^+) and aluminium (Al_3^+) cations in the soil or when base cations such as potassium (K^+), calcium (Ca_2^+), magnesium (Mg_2^+) and sodium (Na^+) are leached and replaced by hydrogen or aluminium (Bolan, Hedley and White, 1991; Helyar and Porter, 1989; von Uexküll and Mutert, 1995). The main causes of soil acidification are: (1) long term rainfall that results in on-site leaching of base cations; (2) draining of potentially acid sulphate soils; (3) acid deposition when urbanization, industrialization, mining, construction or dredging release acid substances into the air or water, causing off-site acidification; (4) excessive application of ammonium-based fertilizers (e.g. ammonium sulphate) as part of intensive agriculture cropping practices; and (5) deforestation and other land use practices that remove all harvested materials, often resulting in a drop of the pH in the topsoil. Only the first of these five causes is a natural phenomenon; all others are human-induced.

In natural ecosystems, soils become more acid with time. Consequently old soils, particularly in humid climates or those developed from acidic rocks, are more weathered and acidic than younger soils or soils of dry climates or those developed from more basic rocks (Helyar and Porter, 1989; von Uexküll and Mutert, 1995). Soil acidification is of the greatest concern in soils that have a low capacity to buffer the decrease in pH and in soils that already have a low pH, such as acid soils in highly weathered tropical areas (Harter, 2007; Johnson, Turner and Kelly, 1982). Soil texture and soil organic matter content play an important role in the buffering capacity of a soil and hence in determining how prone a soil is to acidification (Helyar and Porter, 1989; Steiner *et al.*, 2007). Light sandy soils poor in organic matter are the least buffered against acidification.

Acid sulphate soils contain metal sulphides which, when exposed to oxidation, produce sulphuric acid. Inland, acid sulphate soils form naturally in aquatic ecosystems and also as a consequence of human-induced changes to land use and hydrology. Structures regulating water flow such as dams, weirs and locks prevent flushing of metals, salts and organic matter, and promote the build-up of acid sulphate soils. Acid sulphate soils also form in coastal areas and are common in mangrove forests, saltmarsh, floodplains, and salt- and freshwater wetlands (Lin and Melville, 1994; Pons, van Breemen and Driessen, 1982; Pannier, 1979). Due to the abundance of metal sulphides in rocks, mining activities also foster the formation of acid sulphate soils (Dent, 1986).

The atmospheric deposition of sulphur dioxide (SO_2), nitrogen oxides (NO_x) and ammonia (NH_3) leads to acid deposition. This can affect not only areas near to the urban, industrial and mining sites where the oxides are produced and released into the environment, but also sites located far away (Fanning *et al.*, 2004; Menz and Seip, 2004; Mylona, 1996; Orndorff and Daniels, 2004). The term 'acid deposition' includes both wet and dry (gaseous) precipitation, usually in the form of acid rain or fog. Besides affecting the chemistry of soil and water resources, acid deposition directly harms plants and fish. Acid deposition is currently a major concern in fast-developing countries such as China (Chen, 2007).

Land use and soil management play a crucial role in determining the chemical characteristics of the soil. Intensive farming practices that employ large inputs of nitrogen fertilizers and remove large quantities of products increase soil acidity (Barak *et al.*, 1997; Bolan, Hedley and White, 1991). Indeed, the conversion of ammonium to nitrate releases hydrogen ions (H^+) into the soil solution that can potentially lower the soil pH. This is a problem in soils with low ability to buffer the increase in H^+ such as those poor in lime and negatively charged organic matter and clay. Harvesting has the potential to increase soil acidity by removing base cations from the soil. This is an issue in both agricultural and forested areas wherever large amounts of biomass are removed by crop harvesting and deforestation (Cavelier *et al.*, 1999; von Uexküll and Mutert, 1995).

6.4.2 | Impact of soil acidification

On acid soils (pH < 5.5), crops and pastures suffer from the resulting increased phytotoxicity (Al, Fe, Mn, etc.), from the reduced availability of nutrients, and from decreased microbiological activity (Cronan and Grigal, 1995; Robson and Abbott, 1989; Slattery and Hollier, 2002; Sverdrup and Warfvinge, 1993; Whitfield *et al.*, 2010). Onsite soil acidification reduces net primary productivity and carbon sequestration by accelerating leaching of nutrients such as manganese, calcium, magnesium and potassium, resulting in nutrient deficiencies for plants (Haynes and Swift, 1986). On-site soil acidification is also responsible for the development of subsoil acidity (Tang, 2004), for the breakdown and subsequent loss of clay materials from the soil (Chen, 2007), and for the erosion which results from decreased groundcover (Slattery and Hollier, 2002). Soil acidification also leads to off-site effects such as surface water acidification through sediment losses, and groundwater enrichment of soluble metals. In turn, these processes mobilize heavy metals into water resources and the food chain (Driscoll *et al.*, 2003; Reuss and Johnson, 1986; Schindler *et al.*, 1980; Slattery and Hollier, 2002; Voegelin, Barmettler and Kretzschmar, 2003).

6.4.3 | Responses to soil acidification

Soil acidification is an insidious process. It develops slowly and, if not corrected by lime applications for example, can continue until the soil is irreparably damaged (Edmeades and Ridley, 2003; Liu and Hue, 2001; Slattery and Hollier, 2002). Biological recovery can potentially be improved by an increase in pH and acid-neutralising capacity (ANC) (Marschner and Noble, 2000). Of main concern is subsoil acidity, which is particularly difficult to correct with conventional methods (Farina, Channon and Thibaud, 2000; Liu and Hue, 2001; Hue and Licudine, 1999). Actions to mitigate global warming can reduce the emission of pollutants such as sulphur dioxide (SO_2) which contribute to soil acidification (NADP, 2014; Smith, Pitcher and Wigley, 2001; Vestreng *et al.*, 2007). However, soil response to decreases in acid deposition is slow and acid-affected sites may require many decades to recover (Zhao *et al.*, 2009).

6.4.4 | Global status and trends of soil acidification

Soil acidity is a serious constraint to food production worldwide. Traditionally it has been counteracted by applying lime to the topsoil but little could be done to increase the pH of the subsoil. Programmes to improve soil pH have been undertaken largely in developed countries, which are able to implement soil management plans to preserve soil properties and to bear the cost of lime to buffer soil acidity. However, even in developed counties, for example Australia, there have been cases where subsoil acidity increased due to failures in correcting topsoil acidity. In developing countries the situation is more stark as the use of lime is constrained by poverty. As a result, the farmed area affected by acidification is on the rise (Sumner and Noble, 2003). Soil acidification affects not only agricultural areas but also forests and grasslands.

According to Sumner and Noble (2003), topsoil acidity (pH <5.5) affects around 30 percent of the total ice-free land area of the world, and subsoil acidity affects as much as 75 percent. Figure 6.8 illustrates the pH status worldwide. The most acidic topsoils (pH <3.5) in the world are located in South America in areas where deforestation and intensive agriculture are practiced, and also in river deltas populated by mangroves, for example the Amazon and Orinoco Deltas (Marchand *et al.*, 2006; Moormann, 1963). Elsewhere, the regions with the highest presence of acid soils are: northern and eastern regions of North America; South-East Asia; Central and South Africa; and northern regions of Europe and Eurasia.

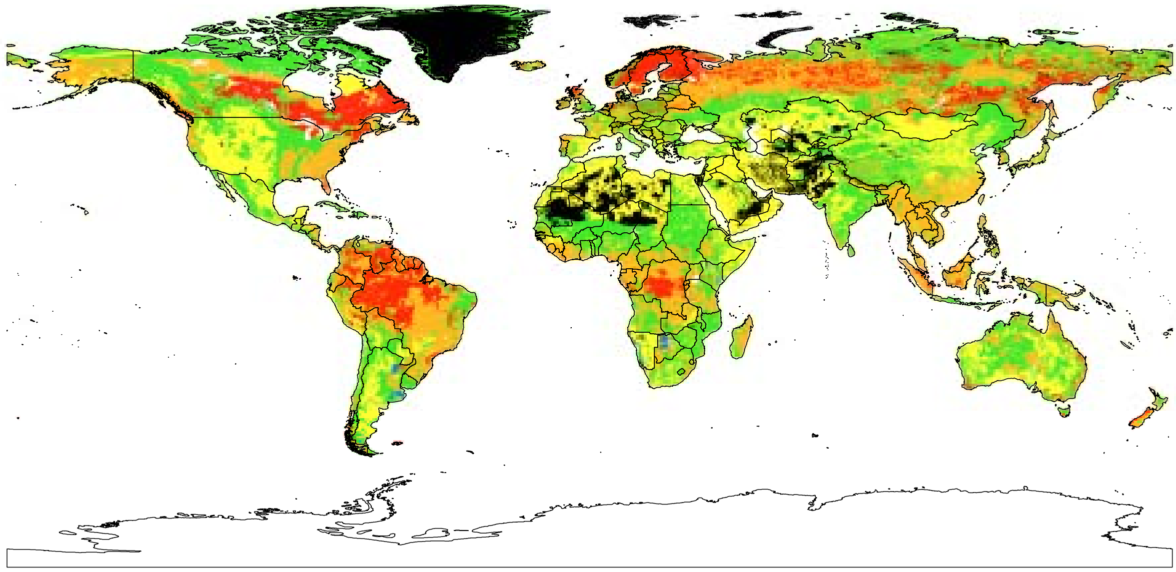

Figure 6.8 | Estimated dominant topsoil pH.
Source: FAO/IIASA/ISRIC/ISS-CAS/JRC, 2009.

The main causes of soil acidification vary by region:

- Regions where soil acidification occurs because of soil texture – parts of North America, Southeast, East and South Asia (Aherne and Posch, 2013; Eswaran *et al.*, 1996; Hicks *et al.* 2008; Shamshuddin *et al.*, 2014; Ouimet *et al.*, 2006)
- Regions where proximity to deltas and coastal plains is a primary cause – parts of West Africa (Bullock *et al.*, 1996),
- Regions where weather conditions are a main cause – parts of Africa and Asia (Breuning-Madsen and Awadzi, 2005; Drees, Manu and Wilding, 1993; Eswaran *et al.*, 1996; Kottek *et al.*, 2006; Wilke, Duke and Jimoh, 1984),
- Regions where acid deposition is an important factor – parts of East Asia and North America (Aherne and Posch, 2013; Quinn, 1989; Wolt, 1981)
- Regions where the massive application of ammonium-based fertilizers plays an important role – parts of East and South Asia (Guo *et al.*, 2010; Wang, Zhang and Zhang, 2010).

In Europe, soil acidification is an issue only in some highly urbanised and industrialized hotspots (EEA, 2010; Kopáček *et al.*, 2004; Menz and Seip, 2004; Moser and Hohensinn, 1983). In the Southwest Pacific, soil acidification is of concern only in intensively farmed areas (Brennan, Bolland and Bowden, 2004; Hartemink, 1998; Xu *et al.*, 2002; Lockwood *et al.*, 2003; NLWRA, 2001). Thus soil acidification affects all regions to some extent, but it is of main concern in poor and developing countries which are growing rapidly but are unable to buffer the decrease in soil pH through conventional means.

6.5 | Global status of soil salinization and sodification

6.5.1 | Status and extent

Salt-affected soils occur in more than 100 countries and their worldwide extent is estimated at about 1 billion ha. Salt-affected soils include those affected by salinity, where the electrical conductivity of the soil exceeds 4dSm–1; and those affected by sodicity, where the exchangeable sodium percentage (ESP) exceeds 6 (Ghassemi, Jakeman and Nix, 1995). Saline soils contain excessive soluble salts, mainly sodium chloride (NaCl) and sodium sulphate (Na_2SO_4) or other neutral salts. These salts increase osmotic pressure, diminish water availability and inhibit plant growth. Sodic soils generally have a low salt content but a high ESP, which causes dispersion of clay particles and results in deterioration of the soil structure. These soils generally have low air and water permeability and a pH above 8.2.

Salinity problems are encountered in all climates and are a consequence of both natural (primary) and human-induced (secondary) processes. Soil salinity and sodicity problems are more common where rainfall is insufficient to leach salts and excess sodium ions out of the rhizosphere. Salt-affected soils often occur on irrigated lands, especially in arid and semiarid regions, where annual rainfall is insufficient to meet the evapotranspiration needs of plants and to provide for leaching of salt. In humid areas, soluble salts are carried down through the soil profile by percolating rainwater and ultimately are transported to sea.

Although salt-affected soils are widespread and an increasingly severe problem, no accurate recent statistics are available on their global extent. The best available estimates suggest that about 412 million ha are affected by salinity and 618 million ha by sodicity (UNEP, 1992), but this figure does not distinguish areas where salinity and sodicity occur together. The Soil Map of the World (FAO/UNESCO, 1980) depicted a similar extent of 953 Mha affected by salinity (352 million ha) and sodicity (580 million ha). Table 6.6 shows the distribution of dryland salinity in different continents.

Human-induced salinity, mainly caused by irrigation without adequate drainage, affects a much smaller area than natural salinity. According to GLASOD, the extent of human-induced salinity is about 76 million ha (Oldeman, Hakkeling and Sombroek, 1991) of which 52.7 million ha occurs in Asia. In Europe, significant parts of Spain and areas in Italy, Hungary, Greece, Portugal, France and Slovakia are also affected by human-induced salinization.

Table 6.6 | Distribution of salt-affected soils in drylands different continents of the world. Source: UNEP, 1992.

Continent	Saline soils (million ha)	Sodic soils (million ha)	Total (million ha)
Africa	122.9	86.7	209.6
South Asia	82.3	1.8	84.1
North and Central Asia	91.5	120.2	211.7
Southeast Asia	20.0	-	20.0
South America	69.5	59.8	129.3
North America	6.2	9.6	15.8
Mexico/Central America	2.0	-	2.0
Australasia	17.6	340.0	357.6
World total	**412.0**	**618.0**	**1030**

In 2006 the global area equipped for irrigation stood at 301 million ha. At present in developing countries, irrigated agriculture covers about one fifth of all arable land, but accounts for nearly half of all crop production and 60 percent of cereal production. About 70 percent of the world area equipped for irrigation is in Asia where it accounts for 39 percent of the cultivated area. India and China each have 62 million ha equipped for irrigation (FAO, 2011). An estimated 60 million ha (or 20 percent of the total irrigated area) are affected by soil salinity, of which 35 million ha are located in four countries e.g. Pakistan (3.2 million ha), India (20 million ha), China (7 million ha) and the United States (5.2 million ha). Other countries with large amounts of salt-affected lands in irrigation districts include Afghanistan, Egypt, Iraq, Kazakhstan, Turkmenistan, Mexico, Syria and Turkey (Squires and Glenn, 2011).

Australiasia has the largest extent of naturally sodic soils of any continent (Table 6.6).

6.5.2 | Causes of soil salinity

The distribution of salt-affected soils varies geographically with climate, landscape type, agricultural activities, irrigation methods and policies related to land management.

Natural causes of salinity and sources of salt

1. *Rock weathering*: Significant quantities of sodium, and to a lesser extent chloride, occur widely in the parent rocks from which soils form. Over time, rock weathering can lead to appreciable salt accumulation in soils if leaching is restricted. Rock weathering is the primary source of salt in seawater.
2. *Sea water and accession of salt in marine sediments*: Saline soils can form from sediments and parent materials that were once under the sea. Likewise, the salts can be due to tidal inundation. Typical examples include the pseudo-delta of Senegal and the Gambia and in the Philippines where coastal tideland reclamation has created about 0.4 million ha of agricultural salt-affected soils. In the United Arab Emirates, areas along the coastal sabkha (salt marshes or lagoonal deposits) are highly salinized (28.8 dS m^{-1}). In the coastal region of the Abu Dhabi Emirate, salinity is more than 200 dS m^{-1} (Abdelfattah and Shahid, 2007)
3. *Atmospheric deposition*: Salt derived from the sea, either deposited via rain or dry fallout, is the primary source of salt across large areas: for example, many millions of hectares in southern Australia. In arid areas, salt can also be derived from dry lake beds and then blown considerable distances by wind (e.g. Eurasia and parts of Australia).

Human-induced causes

1. The management of land and water resources is responsible for the development of human-induced saline and sodic soils. The main causes are:
2. Poor drainage facilities which induce a rise of the groundwater table. This is a major cause of soil salinization in India, Pakistan, China, Kenya and the Central Asian countries.
3. The use of brackish groundwater for irrigation. This is a major cause of secondary salinization in parts of Asia, Europe and Africa.
4. The intrusion of seawater in coastal areas, for example in Bangladesh.
5. Poor on-farm water management and cultural practices in irrigated agriculture.
6. Continuous irrigation over very long periods, particularly in the Middle East.
7. Replacement of deep rooted perennial vegetation with shallower rooted annual crops and pastures that use less water leading to the rise of saline groundwater, for example southern Australia.

6.5.4 | Trends and impacts

Soil salinity is becoming a significant problem worldwide. From the very scattered information on the extent and characteristics of salt-affected soils, salinity and sodicity are rapidly increasing in many regions, both in irrigated and non-irrigated areas. Increasing soil salinity problems are taking an estimated 0.3 to 1.5 million ha of farmland out of production each year and decreasing the production potential of another 20 to 46 million ha. The annual cost of salt-induced land degradation was estimated in 1990 at US$ 264 ha^{-1}. By 2013, the inflation-adjusted cost of salt-induced land degradation was reported as US$ 441 ha^{-1} (Qadir *et al.*, 2014).

6.5.5 | Responses

There are many available responses to contain the salinity threat. These include: (1) direct leaching of salts; (2) planting salt tolerant varieties; (3) domestication of native wild halophytes for use in agro-pastoral systems; (4) phytoremediation (bioremediation); (5) chemical amelioration; and (6) the use of organic amendments.

In several Asian countries, a blend of engineering, reclamation and biological approaches has been adopted to address salinity and waterlogging problems. In Pakistan, engineering solutions included large-scale Salinity Control and Reclamation Projects (SCARPs), which covered 8 million ha at an estimated cost of US$2 billion (Qureshi *et al.*, 2008). Two big drainage water disposal projects were also undertaken. Measures to address the saline soil problem included leaching of salts by excess irrigation, use of chemicals (such as gypsum and acids), the addition of organic matter, and biological measures such as salt-tolerant plants, grasses, and shrubs.

Improvements in on-farm water and crop management have also been practiced. In North America, changes in land use and management practices have reduced the risk of salinization and helped to improve soil health and agri-environmental sustainability.

In Iraq and Egypt, surface and subsurface drainage systems have been installed to control rising water tables and arrest soil salinity. In Iran, Syria and other Gulf countries, crop-based management, and fertilizers are used to combat salinization (Qadir, Qureshi and Cheraghi, 2007). In Iran, Haloxylon aphyllum, Haloxylon persicum, Petropyrum euphratica and Tamarix aphylla are potential species for saline environments (Djavanshir, Dasmalchi and Emararty, 1996). Also in Iran, Atriplex has been shown to be a potential fodder shrub in the arid lands which could bring annual income as high as US$ 200 ha^{-1} (Koocheki, 2000; Nejad and Koocheki, 2000). Breeding of salt tolerant crop varieties (e.g. wheat, barley, alfalfa, sorghum etc.) is also a recognized management response for saline environments. However, most results have been obtained in controlled environments, with few real field results so far.

The use of organic amendments in Egypt showed that the mixed application of farmyard manure and gypsum (1:1) significantly reduces soil salinity and sodicity (Abd Elrahman *et al.*, 2012). Recently, phytoremediation or plant based reclamation has been introduced in the Near East region. In Sudan good responses for control of sodicity have been obtained through phytoremediation. The production of H^+ proton in the rhizosphere during N-fixation from legumes such as the hyacinth bean (*Dolichos lablab L.*) removed as much Na^+ as gypsum application. This indicates the importance of this technology in calcite dissolution of calcareous salt affected soils (Mubarak and Nortcliff, 2010).

6.6 | Soil biodiversity status and trends

6.6.1 | Introduction

Over the last few decades the importance of soil biota for terrestrial functioning and ecosystem services has emerged as an important focus for soil science research. Current evidence shows that soil biota constitute an important living community in the soil system, providing a wide range of essential services for the sustainable functioning of global terrestrial ecosystems and thereby impacting human wellbeing, directly and indirectly (van der Putten *et al.*, 2004). Soil organisms (e.g. bacteria, fungi, protozoa, insects, worms, other invertebrates and mammals) shape the metabolic capacity of terrestrial ecosystems and many soil functions. Below-ground biodiversity represents one of the largest reservoirs of biodiversity on earth (Bardgett and van der Putten, 2014). Essential services provided by soil biota include: regulating nutrient cycles; controlling the dynamics of soil organic matter; supporting soil carbon sequestration; regulating greenhouse gas emissions; modifying soil physical structure and soil water regimes; enhancing the amount and efficiency of nutrient acquisition by vegetation through symbiotic associations and nitrogen fixation by bacteria; and influencing plant and animal health through the interaction of pathogens and pests with their natural predators and parasites.

Fungi and bacteria are important decomposers in the soil. They are remarkably efficient. The smaller the pieces to be decomposed, the faster these microorganisms are able to do their job. Organic waste such as leaf matter and the droppings of herbivores first feed a host of small animals including insects, earthworms and other small invertebrates which live in the plant litter. The combined fauna break up the organic matter, digesting part of it, and thus facilitating the task of the microorganisms and invertebrates that complete the process of decomposition. In turn, soil macro-fauna affect soil organic matter dynamics through organic matter incorporation, decomposition and the formation of stable aggregates that protect organic matter against rapid decomposition. Successive decomposition of dead material and modified organic matter results in the formation of a more complex organic matter called humus, which affects soil properties by increasing soil aggregation and aggregate stability, increasing the cation-exchange capacity (the ability to attract and retain nutrients), and increasing the availability of N, P and other nutrients.

Many scientists have reported the role of macro-fauna in the accumulation of soil organic matter. For example the work by Snyder, Baas and Hendrix (2009), showed that millipedes and earthworms, both by themselves and taken together, reduce particulate organic matter. In addition, earthworms create significant shifts in soil aggregates from the 2000–250 and 250–53 µm fractions to the > 2000 µm size class. Earthworm-induced soil aggregation was lessened in the 0-2 cm layer in the presence of millipedes. Further, Hoeksema, Lussenhop and Teeri (2000) found that in high-N soil with twice-ambient CO_2 there was a higher density of predator/omnivores, lower diversity, and a larger value of Bonger's Maturity Index compared to ambient CO_2. In this experiment, fine root biomass and turnover were significantly greater under elevated CO_2. This indicates higher vigour in plant root development and growth and hence increased carbon sequestration conditioned by enhanced soil biota activity.

Studies also show the role of soil biota (including fungi, bacteria and plant parasitic nematodes) as pathogens and parasites or herbivores in decreasing root and plant productivity or reducing fruit quality. Recent research has focussed on the use of nematode and fungal resistant plant species or of other soil organisms as suppressive agents to modify the pathogens.

6.6.2 | Soil biota and land use

Losses in soil biodiversity have been demonstrated to affect multiple ecosystem functions including plant diversity, decomposition, nutrient retention and nutrient cycling (Wagg *et al.*, 2014). Links between above-ground and below-ground communities (Wardle *et al.*, 2004; De Deyn and van der Putten, 2005; Bardgett and van der Putten, 2014) suggest that factors affecting above-ground extinction may also be affecting soil organisms.

Agricultural intensification, in particular, may reduce soil biodiversity, leading to decreased food-web complexity and fewer functional groups (Tsiafouli *et al.*, 2015). Other driving forces that influence biodiversity in agricultural soils include the influence of crops/plants, fertilizers and pH, tillage practices, crop residue retention, pesticides, herbicides and pollution (Breure *et al.*, 2004; Bardgett and van der Putten, 2014). Soil biological and physical properties (e.g. temperature, pH, and water-holding characteristics) and microhabitat are altered when natural habitat is converted to agricultural production (Crossley *et al.*, 1992; Bardgett and van der Putten, 2014). Changes in these soil properties may be reflected in the distribution and diversity of soil meso fauna. Organisms adapted to high levels of physical disturbance become dominant within agricultural communities, thereby reducing richness and diversity of soil fauna (Paoletti *et al.*, 1993).

The management practices used in many agro-ecosystems (e.g. monocultures, extensive use of tillage, chemical inputs) degrade the fragile web of community interactions between pests and their natural enemies. The intensification of agricultural management may result in increased incidence of pests and diseases, with numerous studies reporting declines in the biodiversity of soil fauna (Decaens and Jimenez, 2002;

Eggleton *et al.*, 2002). In addition, the contribution of soil fauna globally to organic matter decomposition rates may be highly dependent on the temperature and moisture of an ecosystem (Wall *et al.*, 2008). This underlines the need for global-scale assessments. In a global study of soil fungi using 365 soil samples from natural ecosystems, Tedersoo *et al.* (2014) found that distance from the equator and annual precipitation had considerable effect on fungal species richness. They also identified various other controls on soil fungi and this is starting to provide a benchmark for assessing the impacts of human activities on an important component of soil biodiversity.

Soil management strongly influences soil biodiversity, resulting in changes in abundance of individual species. Using a soil biodiversity pressure index calculation from the European Soil Data Centre, Gardi, Jeffery and Saltelli (2013) estimated that 56 percent of soils within the European Union have some degree of threat to soil biodiversity. Based on a questionnaire completed by 20 experts, the study found that the main anthropogenic pressures on soil biodiversity are (in order of importance): (1) intensive human exploitation; (2) reduced soil organic matter; (3) habitat disturbance; (4) soil sealing; (5) soil pollution; (6) land-use change; (7) soil compaction; (8) soil erosion; (9) habitat fragmentation; (10) climate change; (11) invasive species; and (12) GMO pollution (Gardi, Jeffery and Saltelli, 2013).

There is some experimental evidence that there may be threshold levels of soil biodiversity below which functions decline (e.g. Van der Heijden *et al.*, 1998; Liiri *et al.*, 2002; Setälä and McLean, 2004). However, in many instances this is at experimentally prescribed levels of diversity that rarely prevail in nature. Although some studies demonstrate some functional redundancy in soil communities (e.g. Setälä, Berg and Jones, 2005), high biodiversity within trophic groups may be advantageous since the group is likely to function more efficiently under a variety of environmental circumstances, due to an inherently wider potential. In a synthesis of diversity-function relationships of soil biodiversity focusing on carbon cycling, Nielsen *et al.*, (2011) concluded that although there is considerable functional redundancy in soil communities for general processes, change may readily have an impact on specialized processes. However, data to support this conclusion are still limited. More diverse systems may be more resilient to perturbation since, if a proportion of components are removed or compromised in some way, others that prevail will be able to compensate (Kibblewhite, Ritz and Swift, 2008).

6.6.3 | Conclusions

A comprehensive global-assessment on below-ground biodiversity has yet to be carried out. Although there is a Global Soil Biodiversity Atlas (EU/JRC, in press), no benchmark values exist on a global scale. This makes it difficult to quantify changes or future losses that may result from natural or anthropogenic-induced changes. Although progress is being made, few monitoring programs exist that quantify soil biodiversity across regions and at multiple trophic levels, especially outside of Europe. Regarding the threats to soil biodiversity and the effects on ecosystem functioning, more comparative and coordinated studies (from local to global scales) are needed across all ecosystems. These studies should quantify threats and determine the consequences of soil biodiversity loss to ecosystem functions, as well as the effects of interactions between threats. In addition, there is a need for standardization of methods in soil biodiversity studies so that multiple datasets can be synthesized and benchmark values for global soil biodiversity may be established. The use of DNA-based approaches is accelerating the speed at which data is being collected for all organisms. However, although sequencing data must be deposited into a public database (e.g. Genbank) before publication, the majority of morphological data still remains inaccessible and hence largely unavailable for meta-analysis. International initiatives such as the Global Soil Biodiversity Initiative[2], ECOFINDERS, and the EU-sponsored Global Soil Biodiversity Atlas are steps in the right direction but a common database of soil biodiversity data for both morphological and molecular data is still needed (Orgiazzi *et al.*, 2015).

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6.7 | Soil sealing: status and trends

For millennia, the vast majority of people lived a rural life, largely dependent on agriculture and other rural occupations. Only over the last two centuries has the ratio between the urban and non-urban population started to change rapidly. In 1800, only 3 percent of the world's population lived in cities; in 1900 14 percent, 47 percent in 2000, 50 percent in 2007, and 54 percent in 2014. The proportion of the urban population is expected to rise to 66 percent by 2050 (Figure 6.9).

The world's urban population is growing and cities are expanding in order to accommodate the increasing population and economic activity. It is not known with any certainty what share of the Earth's land surface (ca. 144 million km^2) is now occupied by cities or how much land will be required to accommodate the expected urban expansion (Potere *et al.*, 2009). One of the most accurate estimates of the extent of urban areas at global scale, based on the use of MODIS satellite data at a resolution of 500 m, indicates for the year 2000 an area of 657 000 km^2 (Potere *et al.*, 2009), equivalent to 0.45 percent of the Earth's land surface. Urbanization is an important contributor to regional and global environmental change (Foley *et al.*, 2005, Ellis and Ramankutty, 2008). The growth of cities has a vast impact on the landscape and significant impact on soil resources (Chen, 2007; Gardi *et al.*, 2014).

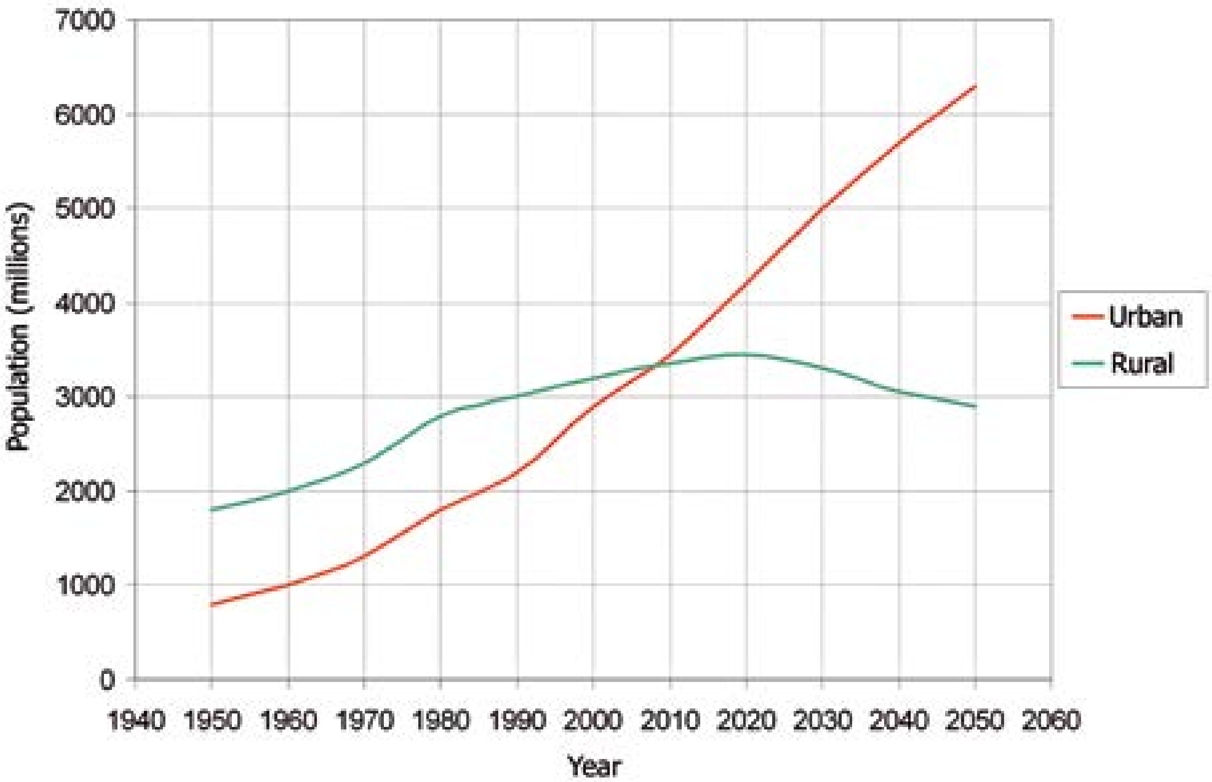

Figure 6.9 | Historical and predicted shift of the urban/rural population ratio.
Source: UN, 2008.

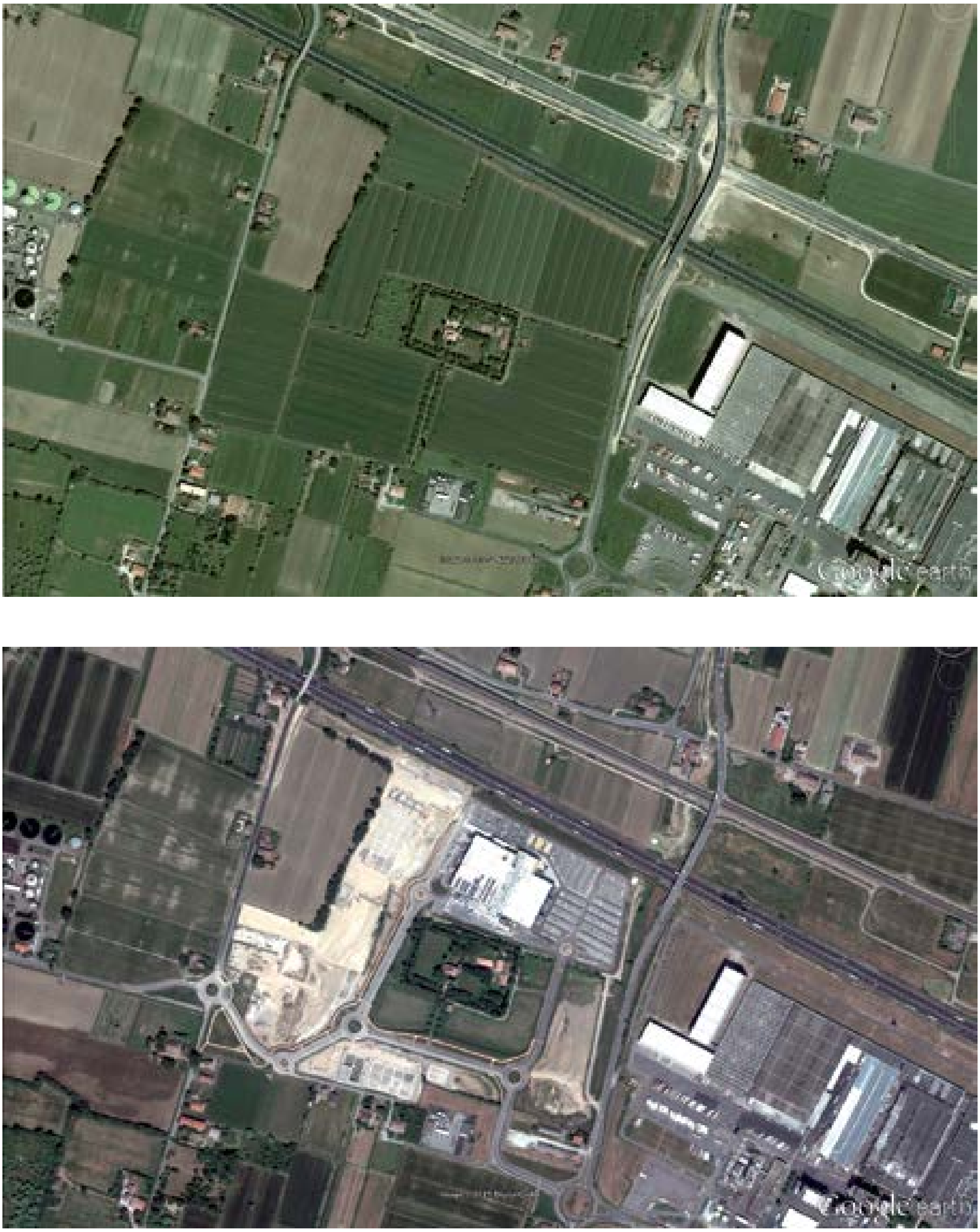

Figure 6.10 | Urbanisation of the best agricultural soils.

Between 1990 and 2000, the total extent of urban areas worldwide increased by 58 000 km^2. During this period, 2.8 percent of Europe's total land was affected by land use change, including a significant increase in urban land. Of the total land take in the EU between 1990 and 2000, 71 percent was for agriculture. Between 2000 and 2006, the equivalent figure was only 53 percent. Had the land taken for urban expansion been devoted to agriculture instead, the land would have produced more than 6 Mt of wheat. More generally, the best quality soil in alluvial plains is often sealed by expanding cities and the rate of conversion is expected to increase rapidly, especially in developing countries.

The term 'soil sealing' is defined as the permanent covering of the soil surface with an impermeable material. Urbanisation affects the inner urban ecosystem as well as the neighbouring ecosystems. Besides the economic and social effects, negative environmental effects are predominantly linked to land consumption, the loss of high quality agricultural soil (Figure 6.10), the destruction of habitat, fragmentation of existing ecosystems, increased fuel consumption, air, water and soil pollution, and the alteration of microclimate.

Soil sealing is in practice equivalent to total soil loss – virtually all services and functions are lost except the carrying capacity as a platform for supporting infrastructure. The main negative impacts on ecosystem services include: virtually total loss of food and fibre production; a significant decrease or total loss of the soil's water retention, neutralization and purification capacities; the loss of the carbon sequestration capacity; and a significant decrease in the ability to provide (micro) climate regulation. The results include the loss of habitat for soil organisms, loss of soil biodiversity and nutrient cycling, and often a diminished landscape and natural heritage.

Urban expansion is, of course, both beneficial and essential. Historically, the beginning of the most important civilizations was associated with both the development of agriculture and the creation of urban settlements. As early as 3000 BC, cities had arisen in the Fertile Crescent, on the banks of Nile, in the Indus River valley and along major rivers in China. However, the very rapid urban expansion of recent times is creating the need for trade-offs, including decisions regarding soil health and the rate of soil sealing.

6.8 | Soil nutrient balance changes: status and trends

6.8.1 | Introduction

Though changes in soil nutrient balances may possibly affect all types of terrestrial ecosystems, rapid changes are more likely to occur in managed ecosystems as a result of the export of biomass or the addition of nutrients to sustain productivity. These managed ecosystems include cropland, intensively or extensively grazed rangelands or meadows, and forests. Monitoring changes in soil nutrient content is of particular relevance in managed ecosystems because it provides a means to evaluate future changes in the ability of soils to maintain their ecosystemic functions. On the one hand, negative balances ('nutrient mining') ultimately translate into crop nutrient deficiencies (, food production deficits and human nutritional imbalances. On the other hand, positive balances may lead to negative environmental and health externalities. Eutrophication, increased frequency and severity of algal blooms, hypoxia and fish kills and loss of habitat and biodiversity have been related to excessive inputs of N and P into fresh and coastal waters. Excess application of N has also lead to widespread contamination of groundwater by NO_3. Gaseous emissions of ammonia and nitrous oxide may also degrade air quality and contribute to acidification, eutrophication, ground-level ozone and climate change (Oenema, 2004; Chadwick *et al.*, 2011). In addition, strongly positive balances may reflect poor economic management of managed ecosystems. Nutrient balances can thus be viewed as indicators of sustainability of human-induced land use changes and land use practices.

Soil nutrients include the macronutrients nitrogen (N), phosphorus (P), potassium (K), calcium (Ca), magnesium (Mg) and sulphur (S). In addition, the soil supplies micronutrients (boron, copper, iron, manganese, chloride, molybdenum, zinc), whose concentrations in plants are typically one or two orders of magnitude less than those of macronutrients. In most cases, N, P and K taken individually or in combination are the most limiting nutrients for plant growth. This section will therefore focus on these three elements. In soils, these nutrients may be present in different pools. Because the amount of nutrients in certain pools may vary strongly and erratically over short time intervals, stocks and mass balances are generally calculated on the basis of total nutrient content, without distinction among different forms (Roy *et al.*, 2003).

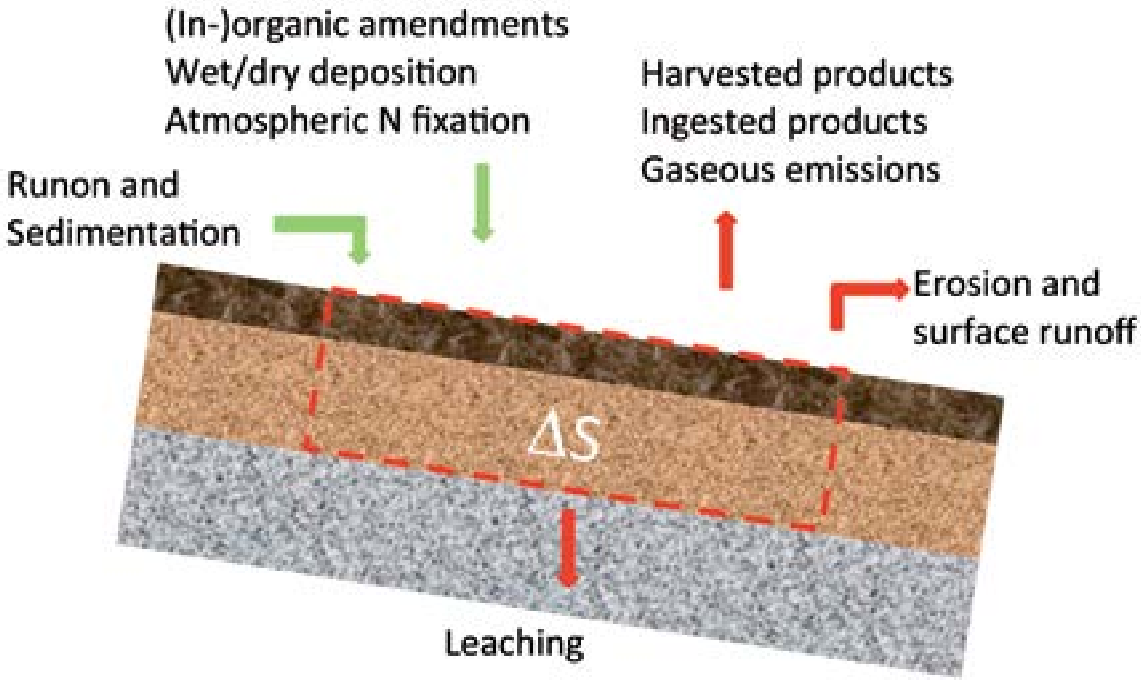

Figure 6.11 | Major components of the soil nutrient balance.

The red discontinuous line marks the soil volume over which the mass balance is calculated. Green arrows correspond to inputs and red arrows to losses. ΔS represents the change in nutrient stock.

6.8.2 | Principles and components of soil nutrient balance calculations

Because the magnitude of the nutrient fluxes is often small compared to the total stock of nutrients in the soil profile, changes in soil nutrient stocks can be rather slow and difficult to detect over short time scales (< decades). Hence calculating nutrient balances from nutrient flows rather than from changes in nutrient stocks has been preferred in many studies (Figure 6.11).

Table 6.6 lists the main inputs and outputs used for calculating the mass balances of N, P and K. Inorganic amendments are mostly composed of mineral fertilizers, but also comprise urine or minerals contained in irrigation water. Organic amendments include liquid, semi-solid or solid manures, compost, mulching material not produced on-site, and household refuse. It also includes faeces dropped by animals. In systems such as urban gardening, the re-use of waste water may also input organic compounds. Biological fixation by bacteria is restricted to N. Wet deposition refers to nutrients supplied with rainwater, whereas dry deposition refers to nutrients deposited as dust and aerosols. Dry deposition is a particularly important phenomenon in the case of K in areas downwind of major dust producing areas (e.g. West Africa;). Sedimentation refers to the deposition of sediment eroded upstream or to sediment deposited during river flooding. Additional fluxes may exist in specific situations (e.g. nutrients in subsurface lateral flows ; supply of NO_3 from groundwater ().

Table 6.7 | Major components of soil nutrient mass balances for N, P and K

	N	P	K
Nutrient inputs			
Inorganic amendments	Yes	Yes	Yes
Organic amendments	Yes	Yes	Yes
Biological fixation	Yes	No	No
Dry or wet deposition	Yes	Yes	Yes
Sedimentation and run-on	Yes	Yes	Yes
Nutrient outputs			
Harvested products	Yes	Yes	Yes
Grazed products	Yes	Yes	Yes
Leaching	Yes	Generally negligible	Low
Gaseous emissions	Yes	No	No
Erosion and runoff	Yes	Yes	Yes

The main losses are related to nutrients contained in exported harvested products (crops or fodder), and nutrients contained in food ingested by primary grazers (Table 6.7). Nutrients may also be lost by gaseous emissions (NH_3, N_2, N_2O), through erosion and in surface runoff, or by leaching. The latter applies mostly to NO_3-N, to a lesser extent to NH_4-N and K, and to a very limited extent to PO_4-P except in coarse textured soils saturated with P.

6.8.3 | Nutrient budgets: a matter of spatial scale

The larger the spatial scale, the more certain nutrient flows are internalized. For instance, in a self-sufficient, well-managed farm, the net balance may be nil or close to nil. However, different parts of the farm may well have very different balances. Likewise, in extensively-managed agropastoral systems, nutrient flows mediated through livestock occur between rangelands and croplands. At a regional scale, the balances may thus be nil or only slightly negative, whereas large imbalances exist within the region (see Box 6.1).

At the global scale, fertilizer use and the growing of leguminous crops have resulted in a doubling of the rate at which N enters the terrestrial ecosystems as compared to pre-industrial levels. Likewise, the use of P fertilizers, animal feed supplements and detergents has led to a doubling of P inputs in the environment as compared to background P release from weathering. This is indicative of a net positive balance but hides large regional disparities. Bouwman, Beusen and Billen (2009) calculated global soil N and P balances for the year 2000. Outputs were restricted to harvested and grazed crops and grasses, whereas inputs included manure, fertilizers, N deposition and N fixation. These authors estimated the inputs to soils at 249 Tg N and 31 Tg P yr^{-1} and losses through harvest and grazing at 93 Tg N and 16 Tg P yr^{-1}. Assuming no build-up of N in the soil, their model predicted that 16 percent (41 Tg yr^{-1}) of the inputs may be lost by erosion and leaching, thereby contributing to a loss in environmental quality. In the case of P, their calculations predicted a net accumulation of P at a rate of 12 Tg yr^{-1} and losses of P through leaching and erosion of 2 Tg yr^{-1}. On a continental scale, considering both natural and agro-ecosystems, balances were always positive and comprised between 8.5 (North Asia) and 35 (Europe) kg N ha^{-1} yr^{-1}, and between 0.22 (Africa) and 5.5 (Europe) kg P ha^{-1} yr^{-1}. Focusing specifically on P and cropland, but restricting the balance calculations to fertilizer and manure inputs and harvest outputs, highlighted large P deficits in South America, northern United States and eastern Europe. Large P surpluses were found in the coastal United States, western and southern Europe, East Asia and southern Brazil.

Within the same continent, large variations in nutrient balances may occur. For 13 African countries, estimated balanced or negative nutrient budgets for N, P and K. At the national level, estimated soil nutrient balances for the year 2000 ranged from -2 to -60 kg N ha^{-1} yr^{-1}, from 0 to -1 kg P ha^{-1} yr^{-1}, and from -2 to -61 kg K ha^{-1} yr^{-1}. A later study at 1 km^2 resolution confirmed the overall negative balances but highlighted larger variability over short distances. The rate of nutrient mining by crops was generally low or moderate, because of low land productivity (low yields), but accumulated over many decennia nutrient depletion may become severe and may be strongly aggravated by soil erosion.

Based on a review of 57 nutrient budget studies related to the African continent, confirmed that N budgets at field and farm scale were largely negative whereas for phosphorus negative balances were reported in only 56 percent of the studies. Going from the continental scale to the plot scale, there was a tendency for the variability in nutrient budgets to increase. This is to be expected, as land uses and management practices in smallholder agriculture in Africa are highly diversified between farms, within farms and even within plots. The study did not find a clear trend in the magnitude of the nutrient budgets from plot to continental scales. This is in contrast to other studies which did report increasingly negative balances as the scale increased.

Box 6.1 | Livestock-related budgets within village territories in Western Niger
(Schlecht et al., 2004)

In the Sahelian zone of West Africa, between 1.5 and 9 kg N ha^{-1} yr^{-1} and between 0.06 and 0.7 kg P ha^{-1} yr^{-1} are taken in by grazing livestock. The quantity varies by location and land use type (rangeland, cropland, fallow). However, up to 95 percent of the nutrients consumed by livestock are recycled through faeces. About 40-50 percent of these faeces end up being spatially concentrated at corralling spots or in farmyards, which represent only a few percent of the total village lands. Though nutrient in- and out-flows related to livestock account for only a small fraction of the nutrient flows in Sahelian crop-livestock systems, livestock thus plays a major role in the spatial redistribution of nutrients. Negative balances occur on rangelands and variable (positive or negative) balances are found in croplands depending on the intensity of application of organic amendments.

At even smaller scales, differences in soil fertility may arise from differential nutrient budgets. Strong gradients in soil fertility have been reported around villages, compounds, trees and shrubs as a result of higher levels of inputs (litter, household refuse, human excreta, manure and urine from resting animals, sedimentation, etc.) near these features. These are referred to as 'fertility rings' or 'fertility islands'.

6.8.4 | Nutrient budgets: a matter of land use system, land use type, managementand household equity

Nutrient balances vary greatly across land use (LU) systems. Intensive growing of industrial crops in Europe is generally characterized by excess inputs of N, despite a recent tendency towards reduced fertilization driven by EU regulations and the economics of fertilizer use. As a result of the decoupling of livestock and land and because livestock are increasingly fed with imported feed, pastures are commonly exposed to excessive applications of manure (e.g. in Normandy in France, and in Denmark and Holland). Regarding P, after decades of excess application of P, there is nowadays a tendency for farmers to reduce their P application rates, or even to stop applying P altogether and to rely only on accumulated soil reserves and P released from soil mineral weathering.

At the other extreme, subsistence farming in developing countries is commonly characterized by negative balances, reflecting nutrient mining (Roy *et al.*, 2003). examined nutrient balances for different land uses in a Kenyan district. N deficits in excess of -100 kg ha^{-1} yr^{-1} were found for maize, sugar cane, and pyrethrum. P deficits in excess of -10 kg ha^{-1} yr^{-1} were found for sugar cane, pyrethrum, and beans, but P excesses occurred in tea and maize-bean plots. Except for coffee, tea and seasonal fallow, K deficits in excess of -50 kg ha^{-1} yr^{-1} occurred in all systems. These observed differences reflect differences in the use of (in-) organic amendments, but also nutrient transfers across LU types. In the case of coffee for instance, mulching is recommended, which is done by using residues from other crops (e.g. bananas) or grasses from fallow land.

In Asia, both strongly positive and strongly negative balances have been reported. K deficits have been reported for rice-based systems across several Asian countries ranging from -25 to -70 kg ha^{-1} yr^{-1}. also reported K deficits in 71 paddy farms in south China, but found N and P surpluses. Based on negative nutrient balances for Bangladesh, Vietnam, Indonesia, Myanmar, the Philippines, and Thailand, and positive balances for Japan, Malaysia and Korea, it has been argued that lower-income countries with large and growing population were more likely to present negative balances whereas higher income countries with stable populations tended to have positive balances. In sub-Saharan Africa, the larger the population density, the more negative the N and P balances.

For similar systems, differences in nutrient balances may also arise from variable access of farmers to external inputs. In the Sudanian zone of west Africa, cultivated plots near hamlets tended to have less negative or more positive balances than plots near larger villages because farmers in hamlets cared better for their crops, earned more income from sales and therefore could invest more in fertilizers. Generally, cultivated plots near hamlets and villages benefit from greater additions of household refuse and human and animal faeces. However, social inequality in access to resources has been found to have an equally large or even larger effect on nutrient balances than distance from the village. For instance, positive N, P and K balances were observed for Fulani cropland because their large herds supply them with abundant manure. Likewise, nutrient budgets ranging from strongly negative to strongly positive were reported for banana-based systems in Tanzania depending on access to cattle and cattle management (Roy *et al.*, 2003). Especially in small-holder agriculture, site-specific management may also induce large fertility gradients over short distances.

(Peri-)urban agriculture is characterized by large excesses in nutrients, especially N. This is commonly driven by the market-oriented nature of this production system, which allows farmers to invest in external inputs. In addition, these systems often rely heavily on the re-use of urban solid waste and waste water. Hence, (peri-) urban production systems exemplify another form of large scale fertility transfer, from rural areas to urban areas. Food produced by nutrient mining in rural areas is consumed in cities, leading to strong soil enrichment of urban soils, especially at urban vegetable production sites (see Box 6.2).

6.8.5 | What does the future hold?

Soil nutrient budgets depend on the local socio-economic conditions but also on market prices of inputs and on policies. In Western Europe for instance, rising prices of fertilizers and the strengthening of environmental policies has led to reductions in N and P inputs into farmland, and this trend is expected to continue. Dwindling P resources and climate change may further affect soil nutrient balances, in managed but also in natural ecosystems.

Box 6.2 | Nutrient balances in urban vegetable production in West African cities

Based on a two year study of urban gardening sites in Niamey (Niger), it was found that N, P and K balances were all positive, with values for high and low input gardens respectively of 1133 and 290 kg ha^{-1} for N; 223 and 125 kg ha^{-1} for P; and 312 and 351 kg ha^{-1} for K. Similar N and P balances were reported for urban vegetable gardens in Kano (Nigeria), Bobo Dioulasso (Burkina Faso) and Sikasso (Mali). However, at these latter sites, K balances tended to be negative. Overall, urban vegetable production sites appear to be major nutrient sinks from which large environmental externalities can be expected.

Bouwman, Beusen and Billen (2009) evaluated the impact of four future development scenarios on nutrient balances for the year 2050. The scenarios, describing contrasting future development in agriculture nutrient use under changing climate, are based on the Millennium Ecosystem Assessment. In the most pessimistic case, the global N balance may increase by 50 percent in the coming decades. In case of proactive policies aiming at closing the nutrient balance, the N balance is expected to remain constant at 150 Tg yr^{-1}. Regarding P, all scenarios predict a future increase in global soil P balance. These global balances hide large variations across regions and even across land uses. Unfertilized rangelands are likely to maintain negative P balances. Scenarios with a reactive approach to environmental problems portray significant increases in N and P balances in Asia, Central and South America and Africa, which can be strongly reduced by a proactive approach. For North America, Europe and Oceania, a shift from reactive to proactive environmental policies could allow limiting the increase in N and P balances, or even a decrease in the overall nutrient balance.

Whereas large positive nutrient balances sustained for extended periods of time in industrialized countries have resulted in negative environmental externalities, positive nutrient balances should not be viewed as necessarily environmentally harmful. Indeed, in many developing regions (e.g. sub-Saharan Africa), positive P balances are needed to restore soil fertility potential depleted by long lasting nutrient mining and to boost the often very low crop yields. Inputs of N in organic form may also be beneficial as part of a strategy to restore the soils' organic carbon stocks. Possible negative environmental externalities should be weighed against the benefits of food security, economic welfare and social well-being. To minimize the negative externalities, the best nutrient management approaches should be promoted through judicious policies.

6.9 | Soil compaction status and trends

Soil compaction is an important problem affecting productivity of soils across the globe. A hidden problem of soils occurring on or below the surface, compaction impairs the function of the subsoil by impeding root penetration and water and gaseous exchanges (McGarry and Sharp, 2003). Soil compaction reduces soil macroporosity e.g. from an optimum of 6 to 17 percent, and hence reduces pasture and crop yield (Drewry, Cameron and Buchan, 2008).

Soil compaction in most circumstances is a function of soil type (texture, mineralogy, organic matter), soil-water content and land management (e.g. tillage practices, traffic, grazing intensity). The problem is not limited to crop land but is also prevalent in rangelands and grazing fields, and even in natural non-disturbed systems. Soil compaction occurs when compressible soils are subjected to traction e.g. in forest harvesting, amenity land use, pipeline installation, land restoration, wildlife trampling (Batey, 2009) or winter grazing (Tracy and Zhang, 2008).

Trampling mechanically disrupts soil aggregates and reduces aggregate stability (Warren *et al.*, 1986) and its effect increases with stocking intensity (Willatt and Pullar, 1983). The degree of damage associated with trampling at a particular site depends on soil type (Van Haveren, 1983), soil water content, seasonal climatic conditions (Warren *et al.*, 1986), and vegetation type (Wood and Blackburn, 1984). Climate is therefore an important determinant of the effects of compaction. Where soil moisture deficits are large, a restriction in root depth may have severe effects but the same level of compaction may have a neglible effect where soil moisture deficits are small (Batey, 2009).

Soil compaction effects are long lasting or even permanent (Håkansson and Lipiec, 2000). Especially in cultivated land, soil compaction is exacerbated by low soil organic matter content. Intensive use of farm machinery including tillage implements such as the mould board, disc ploughs and disc harrows contributes to soil compaction, depending on the pattern of load and stress applied and the number of passes. The initial condition of the soil also plays a role, including soil moisture, organic matter content, bulk density, particle size distribution (including high silt content), and aggregate stability (Materechera, 2008; Horn *et al.*, 2005; Imhoff, Da Saliva and Fallow, 2004). Alfisols, a major soil used for crop production in the tropics and covering approximately 4 percent of the African land mass, are particularly vulnerable. They are strongly weathered and inherently of low organic matter and nutrient status, have a weak structure, and are highly susceptible to crusting, compaction and accelerated erosion (Lal, 1987).

Soil compaction decreases soil physical fertility by impairing storage and supply of water and nutrients, and by increasing erosion hazards and the transport of phosphorus and other nutrients out of the farming system. Soil compaction can reduce crop yields by as much as 60 percent (Sidhu and Duiker, 2006). The range of yield effects is variable, and depends partly on the crop. Cotton was found to be more sensitive to soil compaction than were soybeans, corn or Brachiaria brizantha (Busscher, Frederick and Bauer, 2000). Yields of sugarcane (Saccharum officinarum L.) were reduced by 40 percent with sub-surface compaction of a clay soil (Jouve and Oussible, 1979), while in a clay loam soil wheat yields were reduced by 12 to 23 percent (Oussible, Crookstone and Larson, 1992). The compaction effects on yield are greatest when the crop is under stress, such as from drought or an excessively wet growing season (Sidhu and Duiker, 2006). Krmenec (2000) observed stand count reductions of 20 to 30 percent, plant height decreases of up to 50 percent and yield reductions of about 19 percent in compacted compared to non-compacted plots. The study of Voorhees, Nelson and Randall (1986) illustrates that a one-time compaction event can lead to reduced crop yields up to 12 years later. In another study, soil compaction reduced grass yield by up to 20 percent due to N-related stresses (Smith, McTaggart and Tsuruta, 1997; Douglas, Campbell and Crawford, 1998). In addition, the creation of waterlogged zones or of dry zones caused by shallow rooting can deny plants access to deeper reserves of water (Batey and McKenzie, 2006).

Additional consequences include chemical changes, such as the amount of greenhouse gases (nitrous oxide and methane) emitted from or taken up in a soil (Hansen, Maehlum and Bakken, 1993; Ruser *et al.*, 1998), and reduced root growth and consequently lower crop yields. A study by Gray and Pope (1986) showed also that the incidence of Phytophthora root rot in soybeans (Glycine max. L.) was greater with soil compaction. Soil compaction increases the abundance of anaerobic microsites and decreases the proportion of coarse pores, which may favour emissions of both CH_4 and N_2O (Ball, Scott and Parker, 1999a). Only rarely has soil compaction been associated with positive impacts, such as increasing the plant-available water capacity of sandy soils (Rasmussen, 1985) or reducing nitrate leaching (Badalıkova and Hruby, 1998) or benefiting soybean grown in areas prone to iron deficiency chlorosis in wet years (DeJong-Hughes *et al.*, 2001).

6.9.1 | Effect of tillage systems on compaction

While all tillage methods tend to reduce soil bulk density and penetration resistance to the depth of tillage (Erbach *et al.*, 1992), equipment used in modern agriculture causes soil compaction of topsoil and subsoil. Working the soil to avoid compaction requires timing of tillage in relation to soil water moisture content and soil texture (Håkansson and Lipiec, 2000). No-tillage (NT) agriculture is gaining wide acceptance and is among the top options in the portfolio of technologies to reduce tillage costs, conserve soil and water, increase soil organic carbon (SOC) pools, and reduce net CO_2 emissions, which contribute to global warming (Lal *et al.*, 2004). Despite the numerous benefits of NT, there is no consensus yet on its role in alleviating soil compaction: some researchers report increased compaction associated with the practice (Bueno *et al.*, 2006) and others a decrease in compaction (Gregory, Shea and Bakko, 2005). Increasing soil organic matter, as practiced in conservation agriculture, reduces soil compactibility (Thomas, Haszler and Blevins, 1996), but residue availability remains a key challenge, especially in Africa.

6.9.2 | What is the extent of deep soil compaction?

Soil compaction affects mainly topsoils (Balbuena *et al.*, 2000; Flowers and Lal, 1998) but can also affect subsoils at depths >30 cm. Most subsoil compaction occurs when the soil is wet and field equipment weights exceed 10 tons per axle. The average weight and power of vehicles used on farms has approximately tripled since 1966 and maximum wheel loads have risen by a factor of six (Chamen, 2006). While remediation of shallow compaction is possible, for example by ripping and subsoiling, correcting soil compaction at depths below 45 cm is challenging (Batey, 2009; Berli *et al.*, 2004). Both topsoil and subsoil compaction have been acknowledged by the European Union as a serious form of soil degradation, estimated to be responsible for degradation of up to 33 million ha in Europe (Akker and Canarache, 2001). Similar compaction problems have been reported elsewhere, including in Australia, Azerbaijan, Japan, Russia, China, Ethiopia and New Zealand (Hamza and Anderson, 2005). The total amount of compacted soil worldwide has been estimated at approximately 68 million ha or around 4 percent of the total land area (Oldeman, 1992; Soane and Van Ouwerkerk, 1994). Nearly 33 million ha is located in Europe, where the use of heavy machinery is the main cause. Cattle trampling and insufficient cover of the top soil by natural vegetation or crops account for compaction of 18 million ha in Africa, and 10 million ha in Asia (Flowers and Lal, 1998; Hamza and Anderson, 2003). Agricultural mismanagement (80 percent) and overgrazing (16 percent) are the two major causative factors of human induced soil compaction (Oldeman, 1992).

6.9.3 | Solutions to soil compaction problems

Soil compaction, like soil chemical characteristics, should be monitored routinely and corrected as part of soil management (Batey, 2009). Although soil compaction effects on soil biodiversity and related functions and processes depend on several site and soil properties, a threshold of effective bulk density of 1.7 g cm–3 is the maximum above which only negative effects are observed (Beylich *et al.*, 2010). Managing soil compaction can be achieved through appropriate application of some or all of the following techniques: (a) addition and maintenance of adequate amount of soil organic matter to improve and stabilize soil structure (Heuscher, Brandt and Jardine, 2005); (b) guiding, confining and minimizing vehicular traffic to the absolutely essential by reducing the number and frequency of operations, and performing farm operations only when the soil moisture content is below the optimal range for the maximum proctor density (Kroulik *et al.*, 2009); (c) mechanical loosening such as deep ripping (Hamza and Anderson, 2005); and (d) selecting a rotation which includes crops and pasture plants with strong tap roots able to penetrate and break down compacted soils (Hamza and Anderson, 2005). Promoting macrofauna activity can accelerate creation of channels for water infiltration and root growth. Arbuscular mycorrhiza can to some extent alleviate the stress of soil compaction. This effect has been observed on wheat growth following increased root/shoot ratio of wheat under compaction (Miransari *et al.*, 2008). In the long-term, soil compaction can be reduced by natural processes that cause the soil to shrink and swell such as wetting and drying (Shiel, Adey and Lodder, 1988), and freezing and thawing (Miller, 1980).

Soil moisture lower than the plastic limit is desirable for cultivation. Traffic should be avoided or restricted when condition are otherwise. For farmers, a simple test to avoid soil compaction involves squeezing a small lump of soil into a ball and rolling it into a rod about 3 mm in diameter. If a rod can be made easily, the soil is too wet and will compact if it is worked or has animals or machinery on it. If the rod is crumbly the water content should allow traffic and cultivation without compaction. If a rod will not form at all, the soil could be too dry for tillage in a sandy or loamy soil. This test should be run at several points over the full depth of any proposed cultivation.

6.10 | Global soil-water quantity and quality: status, processes and trends

The world relies on its freshwater for ecosystem health and human well-being and prosperity. Yet only 2.5 percent of the world's water is fresh, and of that, 68.7 percent is in the form of ice. Groundwater comprises 30.1 percent of the freshwater, and just 0.4 percent of the world's freshwater is in lakes, rivers and the soil.

6.10.1 | Processes

Soil water comprises only 0.05 percent of the world's store of freshwater. However, the upward and downward fluxes of water and energy through the soil are massive, and they are strongly linked. The flows are upward in the form of water vapour, long-wave radiation and reflected short-wave radiation, and downward in the form of liquid water and short-wave radiation (Figure 6.12). The soil-vegetation system is the first receiver of the rain and energy that fall on our lands. The soil-vegetation system, which encompasses the upper reaches of the groundwater or basement rock to just above the soil-vegetative layer, is the critical zone for controlling terrestrial water quantity and quality.

Rodell *et al*. (2015) estimate the total annual precipitation onto continents to be 116 500±5 100 km^3 yr^{-1} – equivalent to approximately five-times the water stored in the Great Lakes of North America. Sixty percent of this (70 600±5 000 km^3 yr^{-1}) returns to the atmosphere through evapotranspiration. The remaining 40 percent (45 900±4 400 km^3 yr^{-1}) leaves the continents as runoff, with the greatest proportion either running off the surface of the soil or returning to streams via the groundwater flow system after passing through the soil. Thus small changes due to human intervention and climate change that alter these fluxes can have very large impacts on the store of soil water.

The quantity, quality and flow of water over and through soil affect the spatial and temporal availability and usage of water. The quantity of soil water in a particular layer of soil can be determined by the soil-water retention curve, the so-called 'soil-water characteristic' (Figure 6.13). This curve describes the relationship

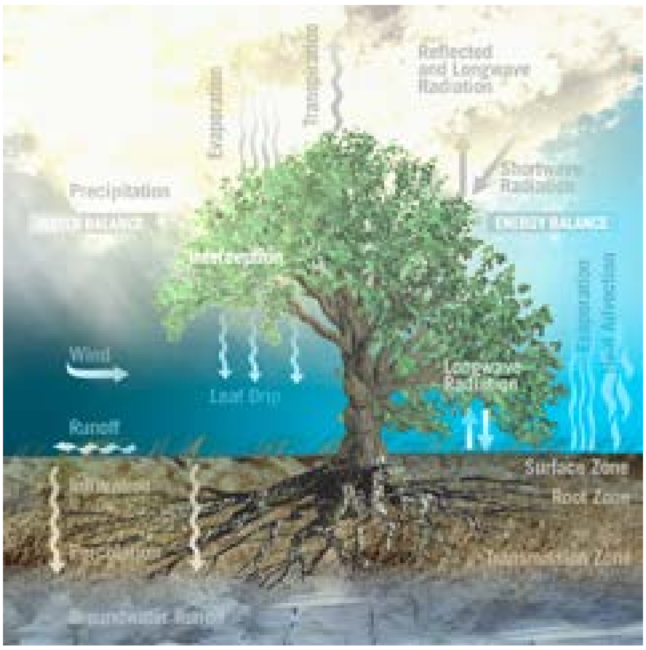

Figure 6.12 | The flows of water and energy through the soil-vegetation horizon

between the amount of water a particular soil can hold and the energy, or matric potential, required to overcome adhesive and cohesive forces to extract water from the soil. Soils of different textures have very differing characteristic curves (Figure 6.13) and this affects the movement and storage of water in the landscape.

The quality of the soil's water is determined by the impurities and pollutants present in the soil water, which may, or may not be adsorbed to and/or exchanged in some part with the soil's reactive matrix materials.

The flow of soil water is determined by the gradient in the matric potential, and the soil's hydraulic conductivity, K (cm day^{-1}) (Figure 6.14), which describes the ease with which water flows through the soil pore space. The hydraulic conductivity curve is highly non-linear and strongly dependent on the soil's water content, θ, and hence matric potential (Figure 6.14). Soil water flow can vary from very slow in soil with small pores, to very fast in soil with large interconnected pores.

The soil-water characteristic (Figure 6.14) is an important factor affecting soil microbiology and rhizosphere ecology. It controls the stability of the spatial and temporal geometry of the soil pore space, which in turn defines the allocation of resources to soil biota, the transport of liquids, gases and solutes to and from roots, and the diversity of microbial habitats (Hinsinger *et al.*, 2009). The soil micro-organisms are largely aquatic in nature and do not inhabit the air-filled pores. They live instead in the liquid phase of the pores, the thickness of which is controlled by the matric potential, which also controls the size and distribution of water-filled pores that provide the hydraulic connectivity through soils.

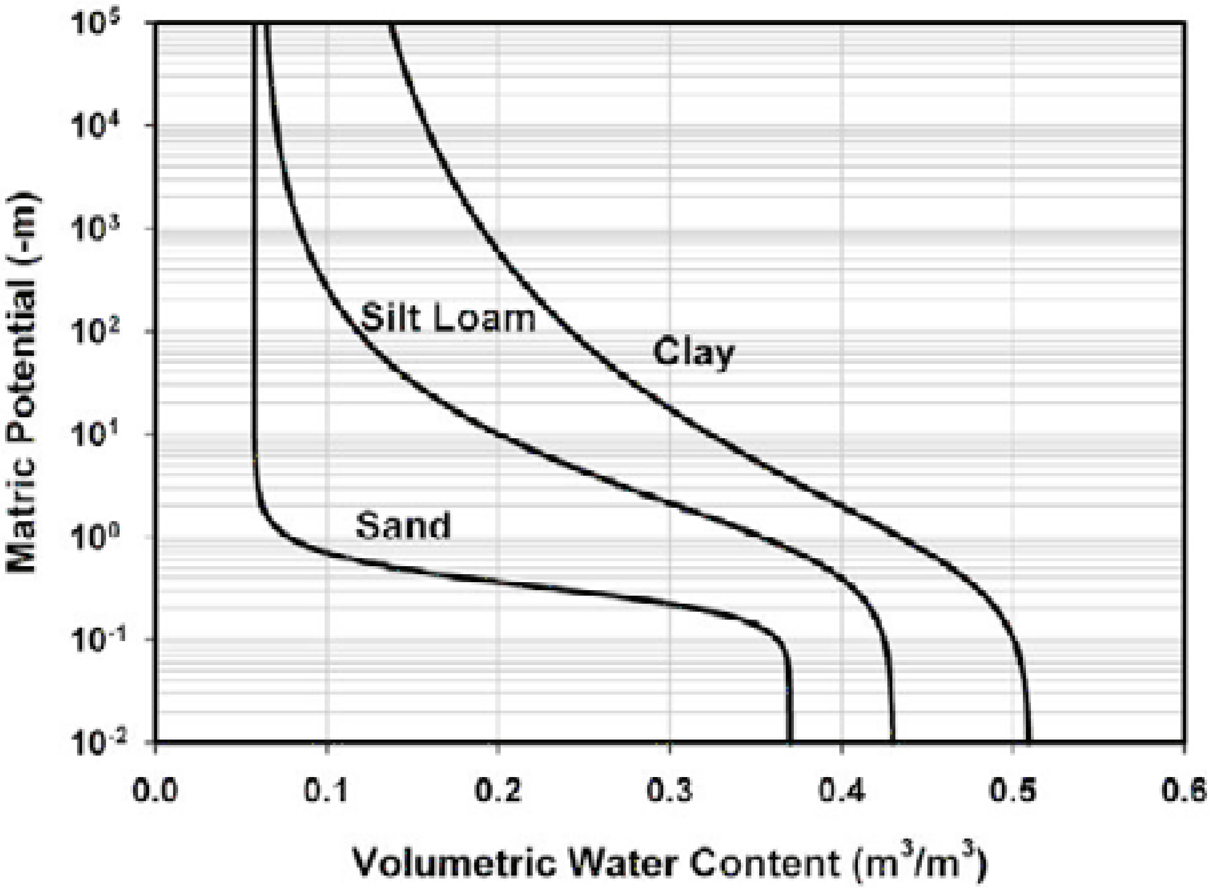

Figure 6.13 | The soil-water characteristic curve linking matric potential, to the soil's volumetric water content.
Source: Tuller and Or, 2003.

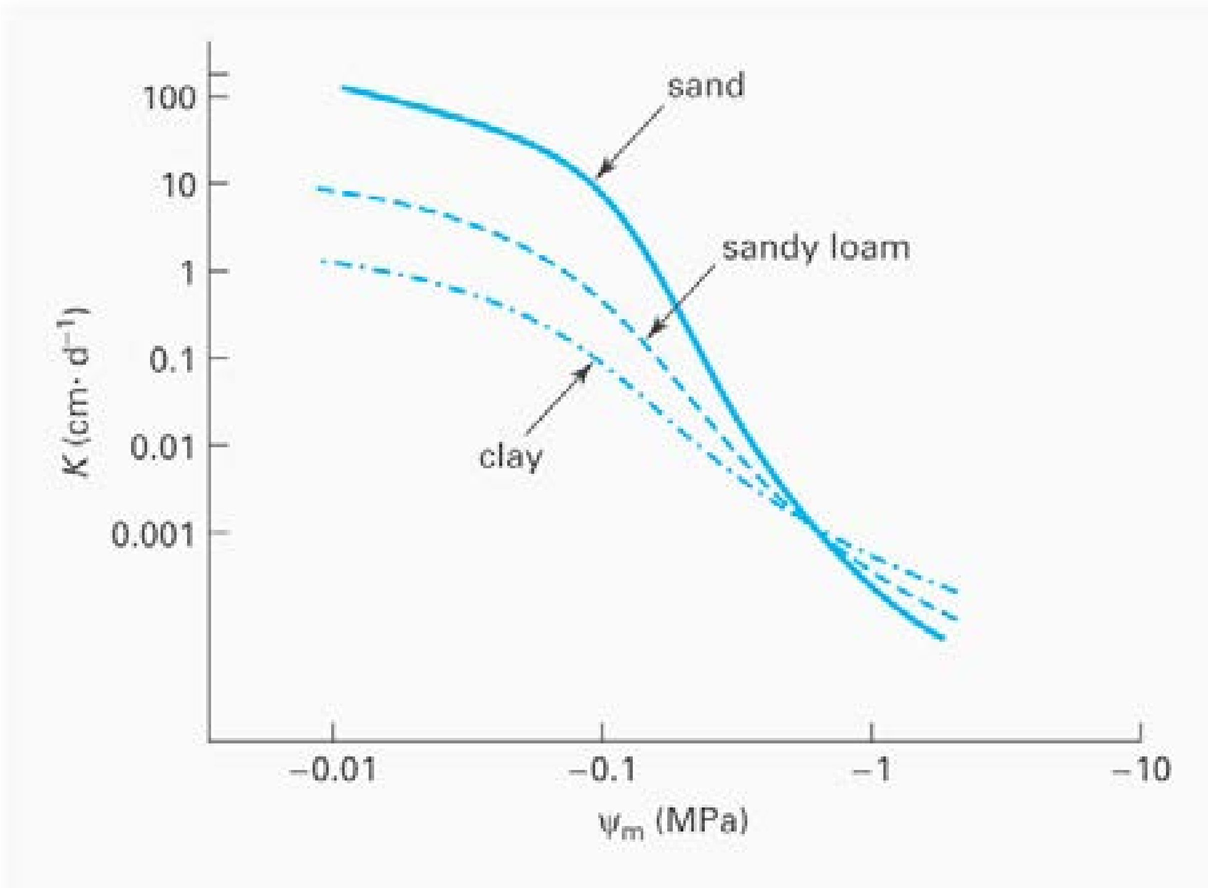

Figure 6.14 | The soil's hydraulic conductivity, K (cm day[-1]) in relation to the matric potential, ψ (MPa). As the matric potential becomes more negative the soil's water content drops (see Figure 6.16) which increases the tortuosity and slows the flow of water.
Source: Hunter College.[3]

The interactions between the structure and physical, chemical and biological components of the soil control the myriad soil functions and processes that are essential for healthy soils, ecosystems and human well-being.

The soil acts as buffer and filter. Indeed, our soil is the world's largest water filter. And through this buffering and filtering, soil controls the quantity and quality of the world's liquid freshwater.

6.10.2 | Quantifying soil moisture

Soil water varies on multiple time and space scales, driven by climate, weather variability, land cover, topography and soil type and structure (Figure 6.15). Measuring variations in soil water is challenging especially at large scales where the cost of direct measurement would be very high. Long-term measurement networks have historically been limited to a few locations globally (Robock *et al.*, 2000). However, with the recognition of soil water as an essential climate variable and the realization that in-situ measurements are necessary for the calibration and validation of remote sensing, the number of operational monitoring networks is increasing (Dorigo *et al.*, 2011). There are also short-term experimental campaigns with multi-scale soil water sampling (Crow *et al.*, 2012). For example, the Soil Climate Analysis Network (SCAN) in the United States provides soil water measurements for 174 sites across the United States, with some measurements dating back to 1992. New technologies such as the COsmic-ray Soil Moisture Observing System (COSMOS) cosmic-ray neutron probes (Zreda *et al.*, 2012) have enabled more efficient and larger measurement footprints of the order of several hundreds of square meters.

3 http://www.geo.hunter.cuny.edu/tbw/soils.veg/lecture.outlines/soils.chap.5/soils_chapter.5.htm

At continental scales, the only practical means of estimating soil water is from satellite sensors or simulation models. Satellite-based measurements of soil water are generally based on measuring microwave emissions that vary because of the sensitivity of the soil dielectric constant to its wetness. These approaches use radiative transfer models to simulate the transfer of radiation emitted from the soil through the vegetation canopy and atmosphere to the satellite sensor. However, measurements have generally been restricted to the top centimetre of the soil column because of the penetration depth of microwave signals for current sensors (> 6 GHz). They are also restricted to sparsely vegetated regions. The recently launched Soil Moisture Ocean Salinity (SMOS) (Kerr *et al.*, 2001) and Soil Moisture Active Passive (SMAP) (Entekhabi *et al.*, 2010) satellite missions improve on this by using L-band (1-2 Ghz) sensors that have penetration depths of the order 5 cm and are less restricted by dense vegetation. Estimates from land surface models have also contributed to understanding the variation of soil water at large scales (Sheffield and Wood, 2008). These simulation models are driven by observations of precipitation, temperature and other meteorology and simulate the surface hydrological cycle with soil water as a prognostic state variable. Recent efforts have developed long-term simulations of soil water at regional to global scales (Sheffield and Wood, 2007, 2008; Haddeland *et al.*, 2011), although uncertainties exist because of missing process representation in the models and because of errors in model structure, parameters and the meteorological forcings.

6.10.3 | Status and trends

Understanding variations in soil water is critical for a range of applications including drought risk management, agricultural decision making, and understanding and attributing climate change impacts. Currently, long-term (multi-decadal) time series of soil water which have been developed from models and satellite retrievals are being used to understand variability and long-term changes in soil water

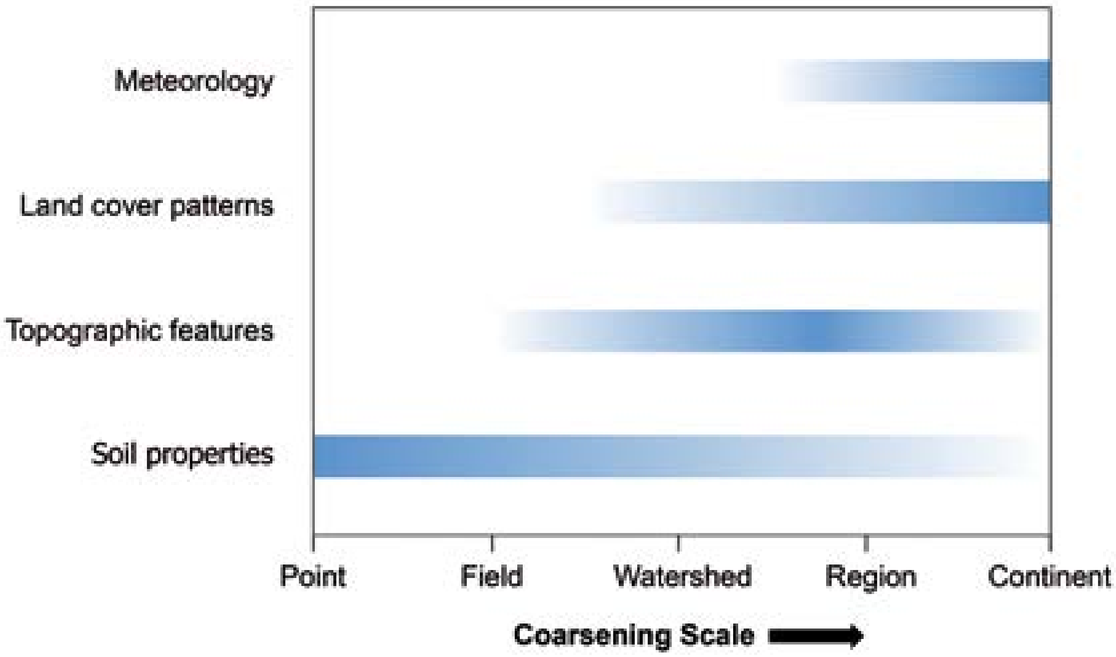

Figure 6.15 | Factors controlling soil water spatial variability and the scales at which they are important.
Source: Crow et al., 2010.

(Sheffield and Wood, 2008; Dorigo *et al.*, 2012). Figure 6.16(a) shows the spatial variability of soil water globally from model simulations, ranging from high values in the wet tropics and northern boreal forests, to the desert regions, such as North Africa, the Middle East, central Asia and Australia. Seasonally, soil water varies with changes in precipitation (Figure 6.16 b) with the largest variations in the monsoonal regions of south and southeast Asia, west and central Africa and the Amazon. From year-to-year, the El Niño Southern Oscillation (ENSO) is the main driver of soil water variability globally (Sheffield and Wood, 2011), often leading to drought conditions in the Amazon, south Asia, eastern Australia and southern Africa during El Niño years, and to drought in the United States southwest and the Horn of Africa in La Niña years.

Longer-term changes in soil water are mostly driven by changes in precipitation (Figure 6.16 c and d). Global warming may be playing a role in drying soil water in some regions, although this is a subject of debate. Over the past 60 years, soil water has been generally wetting over the western hemisphere and drying over the eastern hemisphere, mostly in Africa, East Asia and Europe. Trends over the past 20 years (Figure 6.16 e and f) indicate intensification of drying in northern China and southeast Australia, and switches from wetting to drying across much of North America, and southern South America, in part because of several large-scale and lengthy drought events.

6.10.4 | Hotspots of pressures on soil moisture

Hotspots of pressures on soil water quantity and quality have emerged around the globe. These result from changes in soil water driven by climate change and variability, coupled with human pressures on soil water through, for example, agricultural intensification and extensification. We describe three hotspots: the North China Plain, the Horn of Africa, and the southwestern United States.

The North China Plain has seen rapid expansion of agriculture driven by population growth and increasing demand for food. This area is relatively dry with around 500 mm yr^{-1} of precipitation and so irrigation from groundwater has become an important feature of agricultural intensification. However, groundwater has been used at unsustainable rates, with the result that groundwater levels are dropping by over 1 m per year in some parts (Kendy *et al.*, 2003). Furthermore, precipitation has decreased over the past few decades (Figure 6.16 f). Coupled with intensive irrigation and fertilizer application, this has led to declines in soil water quality through salinization and nitrogen leaching (Kendy *et al.*, 2003).

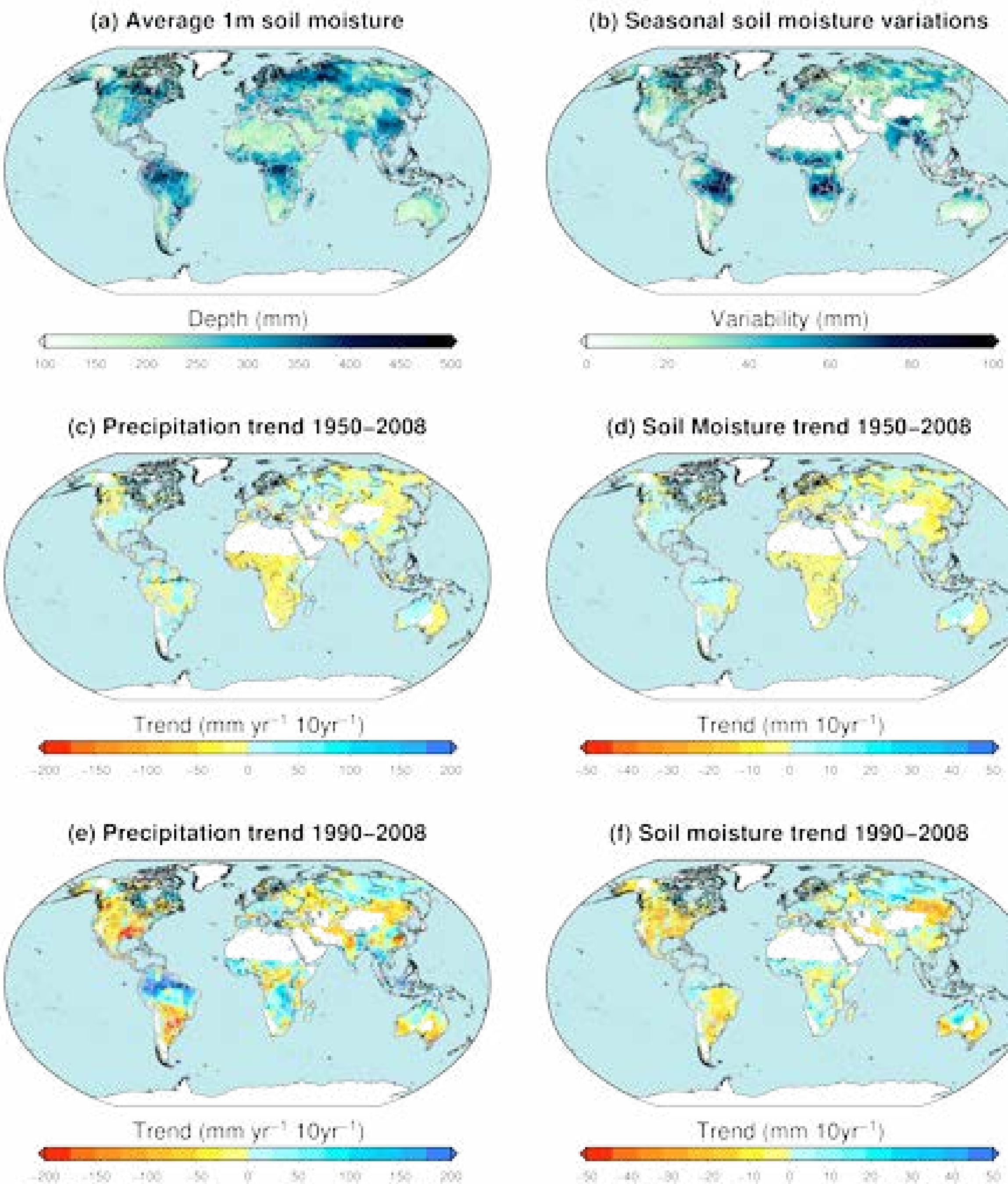

Figure 6.16 | (a) Global distribution of average soil moisture depth in the top 1 m of the soil. (b) Seasonal variability in soil moisture calculated as the standard deviation of monthly mean soil moisture over the year. (c-d) Global trends (1950-2008) in precipitation and 1 m soil moisture. (e-f) As for (c-d) but for 1990-2008. Results for arid regions and permanent ice sheets are not shown. Source: Sheffield and Wood, 2007.

Drought has plagued many parts of Africa because of high climate variability from year to year. Severe droughts in the 1970s and 1980s led to the deaths of hundreds of thousands of people across the Sahel (Sheffield and Wood, 2011). Recent droughts in the Horn of Africa have continued to affect millions of people (Ledwith, 2011; UN, 2011), driven by an overall decline in rainfall that is expected to continue and may be linked to anthropogenic warming of the Indian Ocean (Funk *et al.*, 2008; Williams *et al.*, 2011). Monitoring soil water and its impacts on food security in the Horn of Africa is particularly difficult because of the lack of ground measurements. Nonetheless, the use of satellite and modelling technologies has the potential to provide drought and famine early warning (Anderson *et al.*, 2012; McNally *et al.*, 2013; Sheffield *et al.*, 2014).

Soil water in the southwestern United States has been affected over the past two decades by frequent severe drought events (2000-2002, 2007, 2009), culminating in a three year drought in California (2011-2014) with state-wide impacts on agriculture (Howitt *et al.*, 2014). A shortfall in irrigation water owing to a depleted mountain snowpack was partly offset by increasing groundwater pumping. Recent analysis using Gravity Recover and Climate Experiment (GRACE) satellites has confirmed the resulting massive losses of groundwater since the 1980s from the aquifers underlying California's agriculturally important Central Valley (Famiglietti and Rodell, 2013). McNutt (2014) concludes that "... it is this underground drought we can't see that is enduring, worrisome, and in need of attention".

6.10.5 | Conclusions

Soil water is vital for the health of terrestrial ecosystems and human well-being. Although only a small fraction of the world's water is stored in the soil, the fluxes of water through the soil are massive.

On the time-scale of years, the El Niño Southern Oscillation is the prime control on the global variability in soil water. At longer time-scales, the global pattern of precipitation is the dominant driver in controlling changes in soil water. This pattern may be influenced by climate change.

Global analysis of the changing patterns of soil water has revealed the emergence of three global hotspots in terms of quantity and quality. These are the North China Plan, the Horn of Africa and the southwestern United States. There will be great challenges to address in these hotspot regions and in other pockets where declining soil water quantity and quality is threatening ecosystem health and human well-being.

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