CHAPTER 5

MANAGING THE SOIL

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Soil fertility and climate are the most important factors limiting crop production. The term s*oil fertility* includes the chemical, physical and biological environments in which plant roots grow. Consider the following issues:

- The soil chemical environment provides nutrients to plant roots but undersupply (deficiency) and oversupply (toxicity) are common. Some chemicals not required for plant growth can also accumulate in toxic concentrations. Soils also provide a buffer against adverse conditions such as pH change.
- The soil physical environment is largely controlled by the size and continuity of pores between the soil particles. The system of pore spaces provides pathways through which roots can grow to anchor the plants in the soil and enable them to move through the soil in search of nutrients, water and a favourable gaseous environment. The pore system also affects the ease with which water and gases enter and move through the soil.
- Soils provide the biological environment which contains the plant root system. The surface soil, where organic material accumulates, is the major site for the soil microbial fraction and location for the processes of nutrient cycling. Soil also hosts non-pathogenic as well as pathogenic microorganisms.

The purpose of this chapter is to discuss some of the soil physical, chemical and biological limitations to crop production. Particular focus is on constraints imposed by soil structure, the ability of soil to provide water to plants and the effect of extremes of pH on crop performance. The issues of availability of plant nutrients and the effect of soil borne diseases are important to the performance of crops, they are addressed in Chapter 6 and Chapter 10 respectively.

SOIL STRUCTURE AND STRUCTURAL STABILITY

Soil particles are normally clustered together into *aggregates* or *peds*. The arrangement of the particles within the aggregates controls the proportions of pores of different sizes or the *pore size distribution*. The stability of the aggregates influences how easily the pore size distribution can be changed. It is the pore size distribution within the profile that has a major influence on the ability of a crop to achieve its production potential by controlling:

- the ease by which roots explore soil to access water and nutrients;
- the depth to which roots penetrate;
- the rate of movement of gases (mainly O_2 and CO_2) into and out of soil; and
- the rate of water infiltration and movement to depth as well as redistribution to drying surfaces.

The stability of aggregates also influences the susceptibility of soils to wind and water erosion.

Tillage influences aggregate size and stability, porosity and pore size distribution, and the development of sub-surface layers of high bulk density (hard pans). The passage of machinery over a field influences compaction deeper in the soil profile while the treading by stock compacts surface and sub-surface layers and, during particularly wet periods, causes pugging of the surface.

Hardsetting soils are common throughout Australia. Management of these soils both in terms of chemical treatment and stubble management can markedly influence the performance of crops growing on these soils. To sustain levels of production, it is necessary to manage soil structure. To do this a good understanding is required of the dynamic nature of soil structure and how structure responds to management.

Aggregate Organisation

The organisation of soil particles to form more complex structural units occurs at a series of different scales, ranging from 10^{-7} to 10^{0} m. Kay (1990) illustrated the association of particle size with their respective aggregate forms, the functional role of associated pores of aggregates of particular sized particles and their related biotic phase (Figure 5.1). While this description suggested aggregation of similar sized particles, Dexter (1988) aptly defined soil structure as the spatial heterogeneity of the different components or properties of soil. This definition embraces the reality of soil structure as: *a complex integration of different sized components with different physico-chemical properties existing at different spatial and temporal scales*.

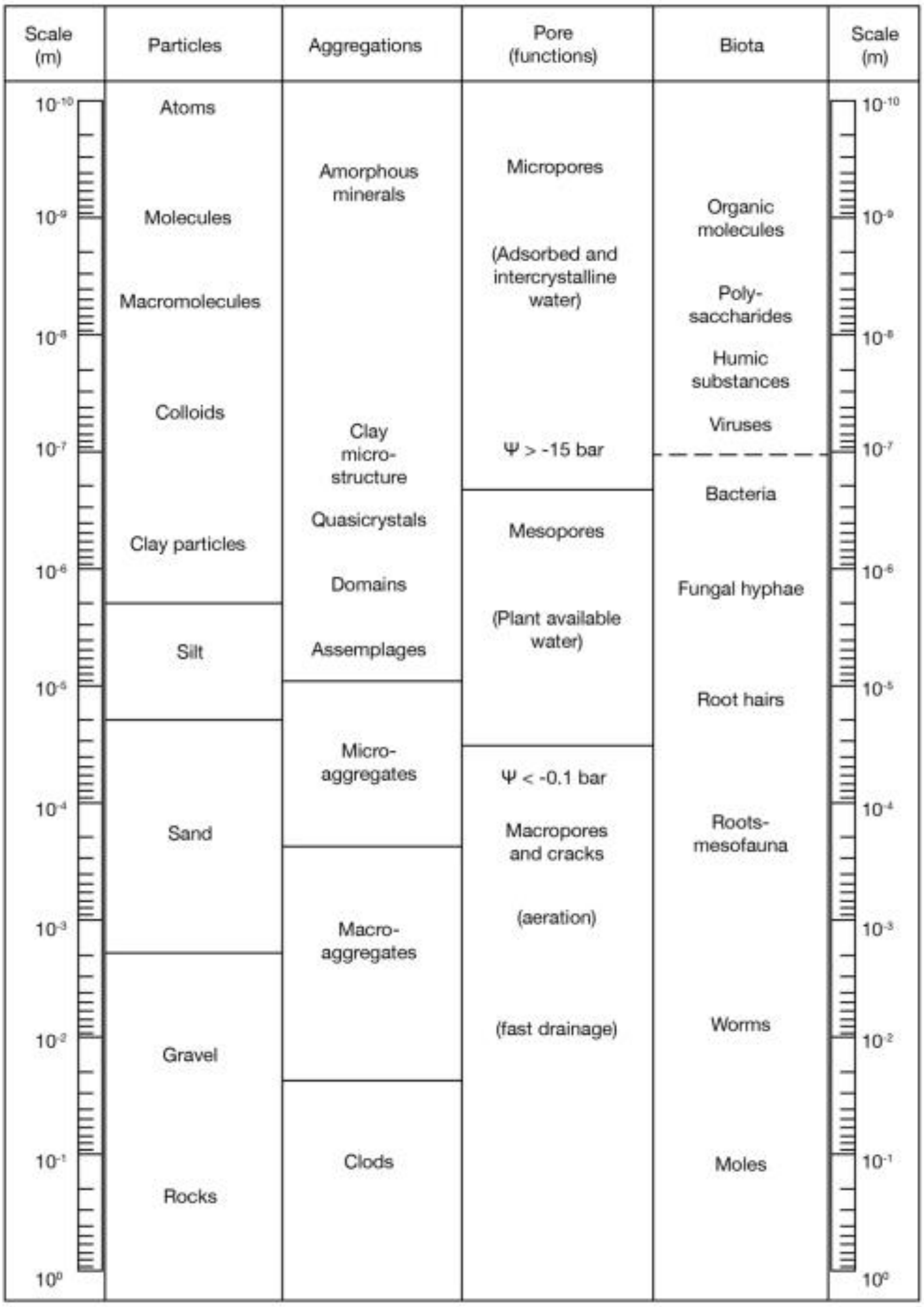

Figure 5.1 Association between particle size and aggregate form as presented by Kay, 1990

Tisdall and Oades (1982) produced a model that illustrated the complexity of aggregates (Figure 5.2). This model relates different aggregate sizes to binding mechanisms. The model suggests that:

- small aggregates (0.2 µm) depend largely on electrostatic binding between opposing clay layers, and the binding of other inorganic solids by inorganic and organic cements; and
- as size and complexity increase, the aggregates of surface soils become increasingly structurally dependent on organic materials such as plant and microbial debris and organic polymers.

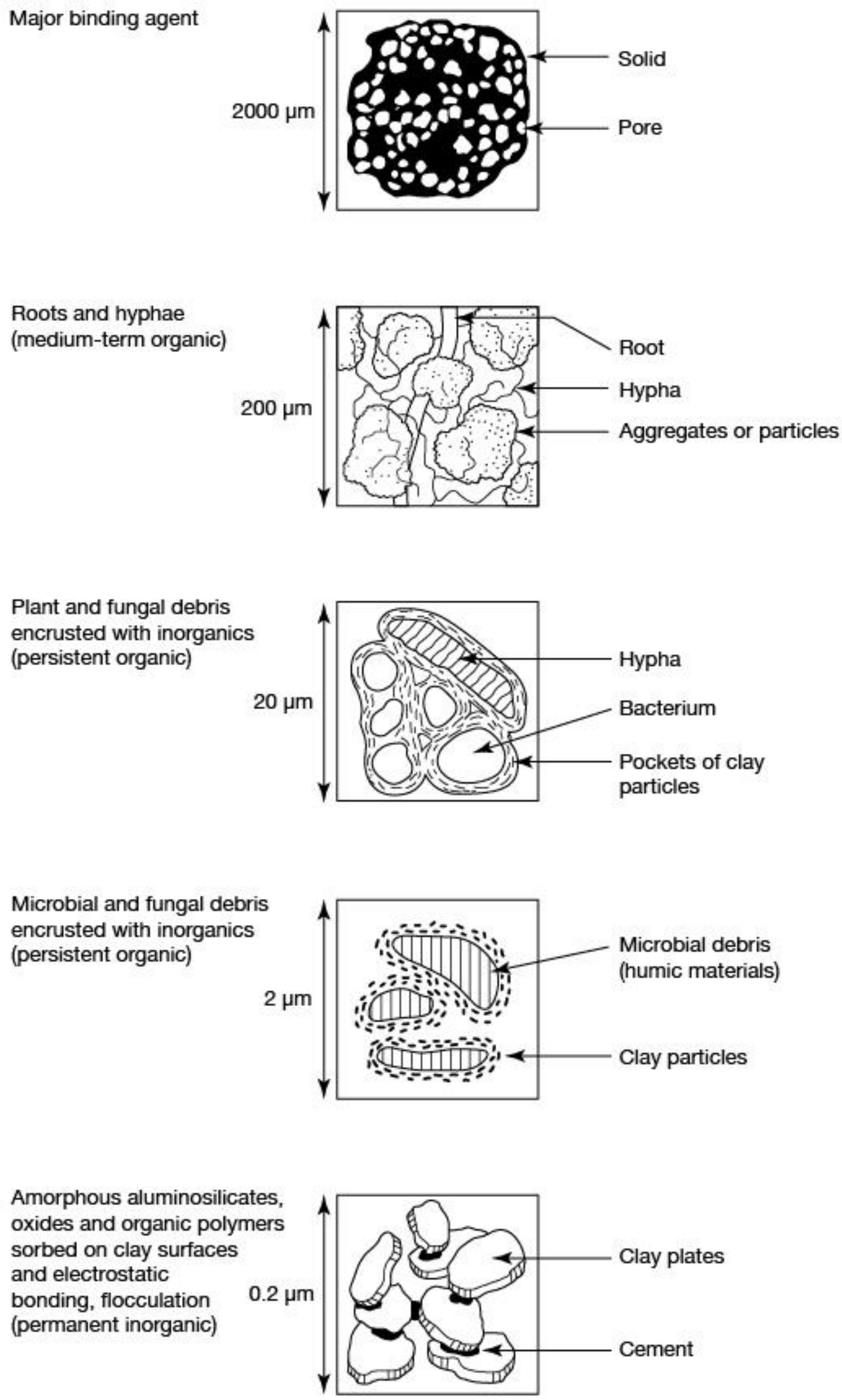

Figure 5.2 Model of aggregate organisation with major binding agents indicated as presented in Tisdall and Oades (1982)

The Development of the Soil Structural Form

Soil structure at any given time is the product of processes of shearing and compression, occurring at different scales and at different rates**.** Deformation by shearing implies a change in shape occurring while maintaining the same volume, while compression implies a change in volume while maintaining shape (Kay, 1990). Intergrades between shearing and compression are more commonly experienced where both the volume and shape of the aggregate change in response to an externally applied force. Shearing and compression caused by the movement of plant roots and soil fauna or by external pressures caused by tillage or livestock, or internal movement caused by wetting-drying or freezing-thawing, can move soil particles in relation to each other, altering the architecture of soil pores and aggregates.

Plant roots affect soil structural form as they pass through existing pores and into the soil matrix creating biopores. As the region behind the root tip expands filling a pore, adjacent pores are compressed. The bulk density in the zone adjacent to the root increases. Organic material released after the death and decay of the root enhances the structural stability of the newly arranged soil particles.

In addition, plants dehydrate soils as they transpire. This water removal can moderate or magnify the structural effects of normal soil wetting-drying events, the effects of which are discussed below.

Soil fauna, such as earthworms, create burrows in soil. The nature of the burrow is dependent on species. Their impact on soil structure is due to the rearrangement of soil particles during movement through soil as well as their contribution to, and redistribution of, organic material.

However, unlike roots, worms cause little lateral pressure as they move through cracks or other points of weakness. Surface litter feeders create largely vertical, semi-permanent burrows that are lined with a mucous-like exudate. Other species, which consume organic material, ingest both organic and inorganic solids, effectively eating their way through soil, creating horizontal pores. The casts remaining in the soil after the passage of earth ingesting worms are a mixture of ingested inorganic and organic material and contribute greatly to the stability of the faunal channel they form.

Drying-wetting of soil imposes differential stresses in the soil, the degree of which depends on:

- the degree of contact of the soil particles to each other; and
- the mineralogy.

The effect of wetting and drying is least pronounced in non-compressive soils, soils whose particles are in full contact with one another at saturation and whose alignment does not change during drying.

However the effect of wetting and drying is prominent in soils containing swelling-shrinking minerals. As these soils dry, shrinkage within the mineral generates a shrinkage tension within the aggregate. When shrinkage exceeds the aggregate's tensile strength, fracturing occurs. The extent of cracking depends on the spatial pattern of water extraction from the soil, the elasticity of the soil and on the state of dehydration of the soil. As a result of shrinkage and cracking, some soil particles are redistributed in relation to each other, creating the potential for an alteration in aggregate architecture upon rewetting.

Upon re-wetting, minerals of fractured soils re-hydrate and swell. The internal swelling of these minerals forces neighbouring particles to realign in relation to one another, breaking some bonds that formed during the previous drying event and altering both the internal architecture and the volume of the aggregate.

The rapid re-wetting of some soils may result in aggregate breakdown via a process known as *aggregate slaking*. Where slaking occurs, the aggregate breaks down into microaggregates. Slaking is common in soils that have insufficient decomposing organic material to provide the aggregate strength to resist internal or externally applied forces.

Freezing-thawing in soil has a significant effect on the architecture of the soil due to:

- the increase in volume of soil water following freezing leading to fracturing of aggregates; and
- the redistribution of soil water in response to freezing. As ice forms in large pores, water moves to the ice and dehydrates the adjacent zone causing localised drying and shrinkage.

The magnitude of the effect of freezing on the soil depends on the size and location of ice crystals formed.

The Structural Role of Soil Organic Material

Organic matter plays a vital role in the binding or stabilising both micro- and macro-aggregates. Tisdall and Oades (1982) proposed the following classification of organic matter with respect to stabilisation of aggregates: *transient, temporary or persistent binding agents.* This classification reflects the chemical nature, age and state of degradation of organic material.

Transient binding agents are short-lived compounds (several weeks) which, while produced rapidly, decompose readily. They include microbial and plant produced polysaccharides. Soil microbes and the growing tips of roots produce polysaccharide rich mucilage and gel that facilitate their movement through soil. These agents are usually effective in improving the transient water stability of aggregates of about 250 μm (Harris *et al*., 1966; Tisdall and Oades 1982).

Temporary binding agents are organic materials that persist in soil for several months to years and comprise mainly fresh and decaying plant roots and fungal hyphae (Tisdall and Oades, 1982). Soil management has a major effect on the production and persistence of these agents. These materials mechanically bind soil particles and smaller aggregates together to form macro-aggregates of diameters exceeding 250 µm. Some examples of the role of this class of organic material are illustrated in electron micrographs presented in Figure 5.3. These images show: (a) the network formed by fungal hyphae around and between soil micro-aggregates and larger soil particles; (b) a remnant root connecting two micro-meso aggregates; (c) the points of attachment of a root to a sand particle; and (d) clay clusters clinging to a fine root or hyphal strand.

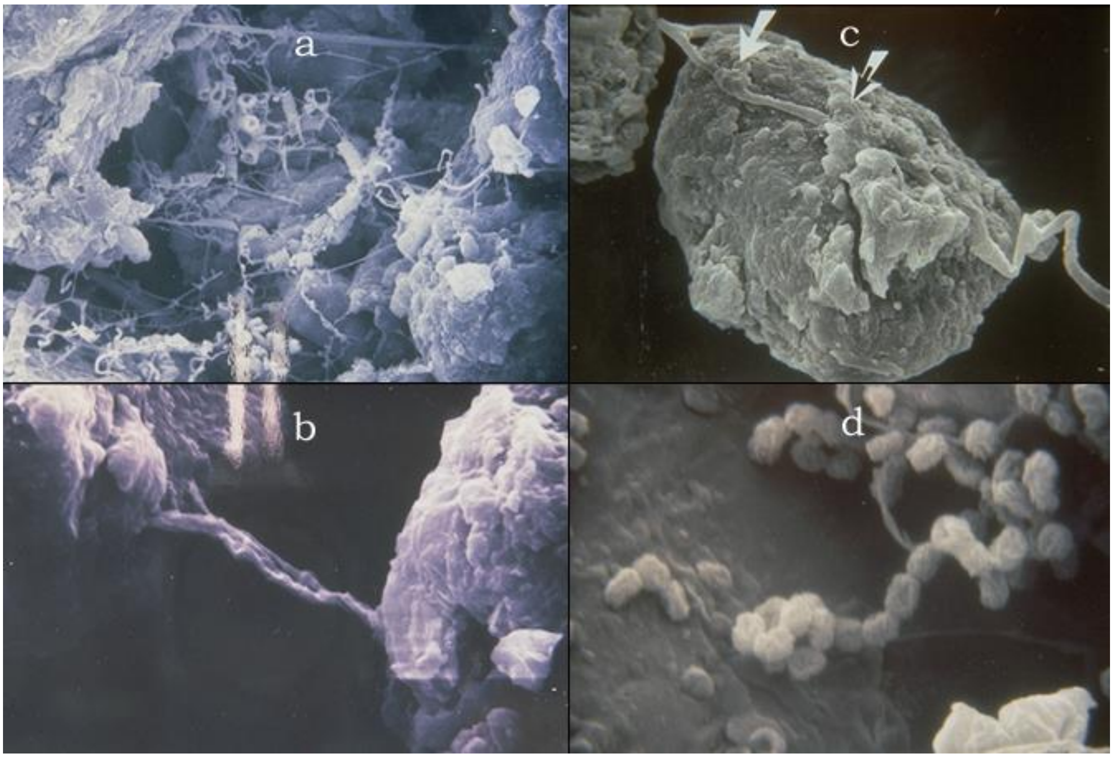

Figure 5.3 Electron micrographs showing various aspects of the structural contribution made by organic material in association with the inorganic fraction of soil: a) network of fungal hyphae connecting microaggregates and larger soil particles; b) a hyphal strand connecting two micro-meso aggregates; c) the points of attachment of a hyphae to a sand particle; d) colloidal amorphous metal-oxyhydroxide clusters or aluminosilicates clinging to a hyphal strand (Scanning Electron-micrographs courtesy of Dr VVSR Gupta)

Persistent binding agents are the remnants of organic matter decomposition generally contained within sterile (pores too small to be accessible by soil bacteria), stable soil pores. These are not affected by management and are likely to be encapsulated by colloidal amorphous oxyhydroxide metals precipitates or aluminosilicates forming larger organo-mineral complexes of the size range 2 – 20 µm (Figure 5.2).

Micro-aggregates (<250 µm) depend heavily on persistent organic materials encapsulated within an inorganic colloidal skin as depicted in Figure 5.2. Turchenek and Oades (1978) showed that most of the organic fraction in the surface layer of a red chromosol (red brown earth) was associated with micro-aggregates (0.4 - 20 µm). The types of organic agents responsible and mechanisms for binding microaggregates have not been clearly identified but are very effective. These aggregates typically contain less organic material per unit weight than macro-aggregates and appear unaffected by agricultural practices (Tisdall and Oades, 1980). It is not unreasonable to assume that there are probably numerous binding mechanisms operable at any one time in micro-aggregates and that the narrow distance between particles is likely to be influential in affecting bond strength.

Macro-aggregates (>250 µm) rely on organic material for binding and structural stability. Research of Tisdall and Oades (1980) showed a strong relationship between water-stable aggregates (>2000 µm) and percentage total organic carbon ($r^2 = 0.93$) (Figure 5.4a). By contrast with micro-aggregates, macro-aggregates were shown to largely depend on decaying plant roots and hyphae for structural stability. In all soil where cultivation was an inclusive part of the rotation, water stability of aggregates (>2000 µm) was related to total root length ($r^2 = 0.81$) and total length of hyphae ($r^2 = 0.77$) (Figure 5.4b,c).

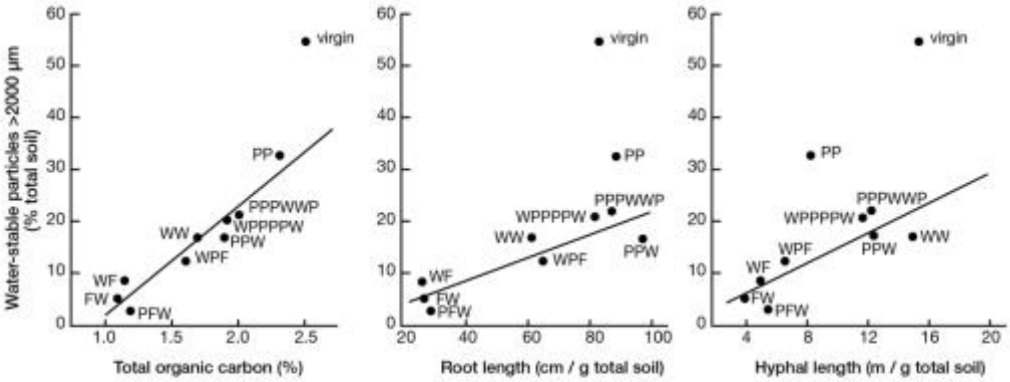

Figure 5.4 Effect of cultural practices, and the relationship between (a) percentage total organic carbon, (b) root length and (c) hyphal length on water-stable particles >2000 µm diameter

(a) Total organic carbon $y = 21.5x - 20.3$ $r^2 = 0.93$ (agricultural soils only):

(b) Root length $y = 0.22x + 0.13$ $r^2 = 0.93$ (cultivated soils only):

(c) Hyphal length $y = 1.45x - 0.02$ $r^2 = 0.77$ (cultivated soils only):
Virgin, never cultivate; pp, permanent pasture; p, annual pasture; w annual wheat crop; f, annual mechanical fallow, complied from Tisdall and Oades (1980)

For most soils, the shape and stability of aggregates are dynamic and show distinct seasonality (Chan *et al*., 1994). The change of season influences the activity of the soil microbial fraction as well as the growth and senescence of plant roots. Maximum soil stability and structural organisation is likely to coincide with periods of maximal soil biological activity (Suwardji and Eberbach, 1998). However, despite the dynamic-seasonal nature of soil structure, agricultural cultivation practices have a major effect on levels of organic matter and, in most soils, strongly influence aggregate stability (Hamblin, 1987).

Compacted and Hardset Soils

Soils compaction refers to the compression of a mass of soil into a smaller volume (Raghavan *et al*., 1990). A soil is considered compacted when the total macroporosity when viewed from an excavation is less than 0.10 m^2m^{-2} (Pagliai, 1988). The effect of compression of the soil volume increases the soil bulk density, decreases total porosity, and increases the proportion of micropores. While hardset and compacted soils cause largely the same symptoms on the affected soil layer, they have different causes. Compacted layers result from the application of external pressures under certain conditions, that is, machinery or stock movements. By contrast, hardset soils compact without the application of external pressure and is discussed in more detail later.

The parameters that best describe compaction, such as porosity or penetration resistance, do not necessarily parallel the response of crops (Raghavan *et al*., 1990).

Compacted soils influence crop production in two ways. Crop performance is directly affected by compaction due to:

- the high mechanical impedance experienced by roots growing in these soils. In contrast with non-compacted soils, the high mechanical impedance of compacted soils restricts root development as well as root density. Raghaven *et al*. (1979) showed that as the number of vehicle passes increased, the depth of root development and the number of roots per gram of soil decreased linearily (Figures 5.5 and 5.6). As well fertilised, irrigated crops are unlikely to suffer much from decreased root proliferation, the effect is likely to be more pronounced in dryland crops where root confinement will restrict access to water and nutrients, particularly in dry years;
- indirectly due to the effect of compaction on porosity and its particular impact on the number of air-filled compared to water-filled pores at field capacity or drained upper limit. The process of compaction compresses macro- and meso-pores to micropore size. The reduction in pore size reduces the rate of entry and movement of water and gas in soils, and decreases the ratio of air to water-filled pores at drained upper limit. In these soils, water movement to the roots of actively transpiring plants is likely to be restricted, causing these plants to wilt due to a temporal lack of soil moisture. During times of heavy rain and low evaporation rates they are likely to be anaerobic.

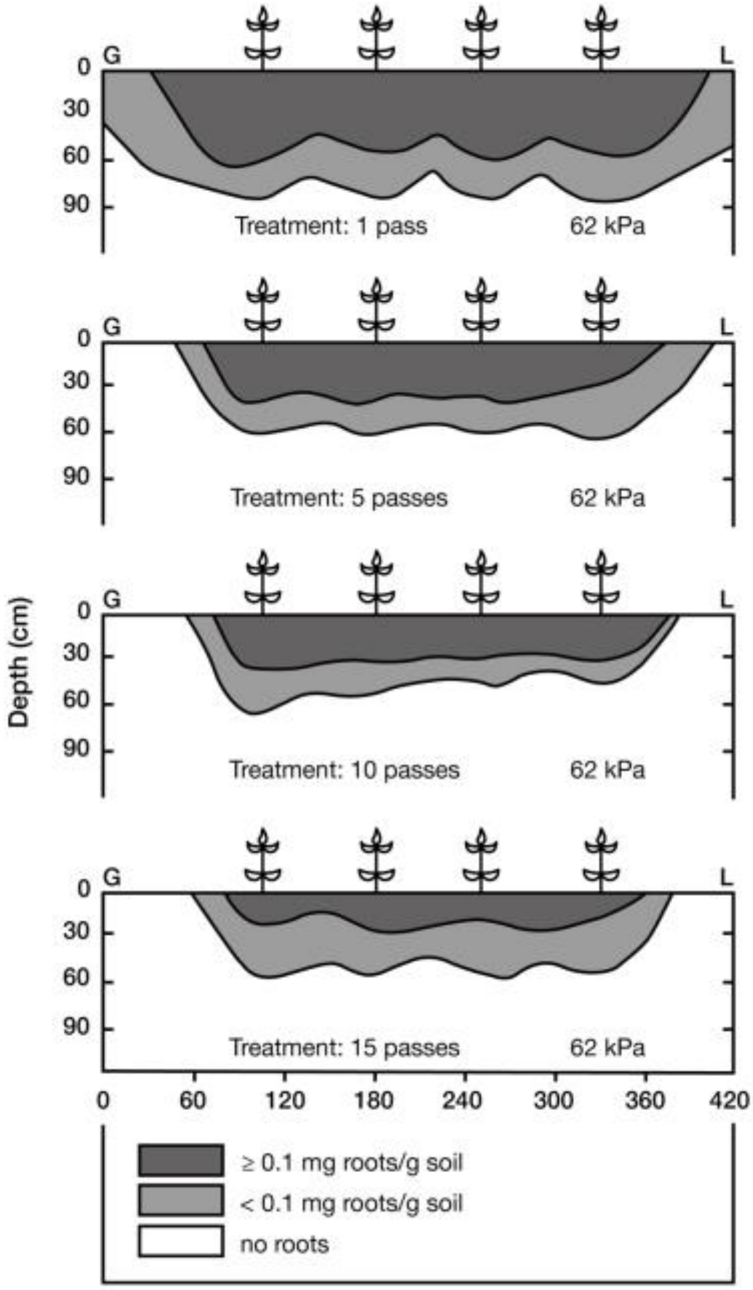

Figure 5.5 The effect of vehicle traffic on the depth of plant root development and on root density (from Raghaven 1990)

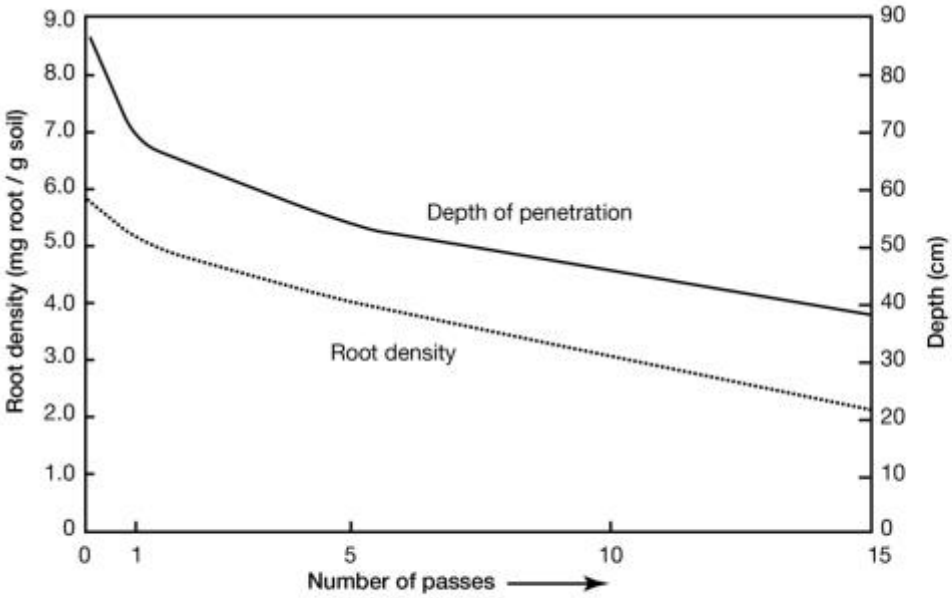

Figure 5.6 The relationship between number of vehicular passes and depth of root development and density (from Raghaven 1990)

Hardsetting soil layers are architecturally similar to compacted soil but are uniquely different in that they form a physically hard, structureless mass upon drying. After this they are difficult to cultivate or cannot be cultivated until rewet (Mullins *et al*., 1990). Hardset soils, like compacted soils, have a layer(s) of high bulk density. However, unlike compacted soil, the hardset layer occurs due to slumping of the soil without the application of external pressure. While approximately 13% of Australian soils have duplex profiles with a hardsetting A_1 horizon, many more have hardsetting crusts or an A_1 with the potential to hardset after cultivation (Mullins *et al*., 1990).

Hardset layers in soil cause similar hydrologic and agronomic problems to compacted soil layers but, in addition, reduce the window during which these soils can be cultivated and reduce the trafficability of these soils when wet.

Hardsetting occurs in soils that have been previously loosened with cultivation where, as a result of wetting, some or all of the aggregates created by cultivation collapse and physically harden when dry, forming a structureless massive horizon. Three processes important in hardsetting are: slumping, uniaxial shrinkage and the development of soil strength.

- *Slumping* occurs as a result of the wetting of soil aggregates that are not water stable. As a generalisation, as water enters dry soil aggregates, energy imparted by the inward movement of water together with the increased pressure of air entrapped within pores, cause the rupturing of bonds which lead to aggregate *slaking.* Following this, some silt or clay particles may *disperse* into solution and the weakened aggregate may collapse under its own weight or slump.
- *Uniaxial shrinkage* occurs as the soils commence drying. During drying, as soil water is depleted, the pores shrink drawing closer together resulting in a

realignment of soil particles. This occurs without forming cracks and, as a result, the soil mass shrinks and bulk density increases.

- *Development of soil strength* As soils start to dry, dispersed soil particles become suspended in the retracting soil water film and lodge in the spaces between non-suspended solids, cementing these particles together. In addition, within small collectives of soil particles, trapped moisture holds particles together via surface tension.

A common characteristic of hardsetting soils is that clays in these soils are generally non-swelling, and dispersive in nature. These clays disperse due to a high proportion of their exchangeable cation base being sodium, that is the soils are sodic.

The international criterion for defining sodicity is an *exchangeable sodium percentage* (ESP) of 15%. In Australian soils, the dispersive influence of sodium may occur at an ESP as low as 6. Australian soils scientists therefore consider soils with an ESP exceeding 6 to be *sodic*. However, while sodium is chiefly responsible in causing dispersion of clays, other soil factors may influence the effect of sodium. These include the presence of:

- multi-valent exchangeable cations with low hydrated radii that increase the ESP at which dispersion occurs. For example Emerson (1983) showed that complete dispersion in one soil subsoil (pH 6.4, ESP 24) contrasted with a lack of dispersion in a more acidic subsoil (pH 5.0, ESP 20);
- higher amounts of organic matter reduce the dispersive effects of sodium. Black and Abdul-Hakim (1985) showed that a soil of moderate sodium concentrations and low salt concentration was less permeable when cultivated than it was under pasture. The difference was attributed to the lower organic matter content of the cultivated soil;
- swelling clays increase the likelihood of dispersion even at low ESP (McNeal and Coleman, 1966) .

Amelioration of hardsetting soils require a number of simultaneous management tools: removal of sodium to prevent clay dispersion; improvement in soil structure to aid sodium leaching and improve the soil hydraulic character; and an increase in the soil total organic carbon content to buffer against the effect of sodium. Gypsum is often prescribed as a chemical amelioration for hardsetting soils as calcium replaces sodium in the diffuse layer and dissolution of gypsum increases the soil solution electrolyte balance, aiding in the flocculation of clays. However improvements in the movement of water in these soils increases the rate of leaching of gypsum, and therefore the impact of gypsum on flocculation is temporary. It is generally accepted that structural improvements from the addition of gypsum lasts up to three years, with further additions likely to be required in time.

The Influence of Structure on Plant Root Development

The soil profile is composed of a series of different horizons, each of which are bio-physically heterogeneous.

In the surface horizon, soil structure and stability are quite variable being the result of soil biological activity, texture, the chemical nature of soils and methods of management.

As depth increases, soil bulk density increases and biological activity decreases. At these lower depths soil structure reflects more strongly the overburden, the small additions of organic carbon, the higher content of clay and the lower density of roots and soil fauna. In these lower layers, soil structure tends toward a blocky form with failure occurring between aggregates in the form of cracks. More extensive cracking may occur in the form of vertical relief in soils containing swelling-shrinking clays.

Movement of plant roots through soil The development and function of plant roots in soil is physically constrained by two boundaries:

- in very hard soils, plant-root elongation may be constrained due to high soil strength, therefore reducing nutrient and water uptake by the plant;
- in loose soil, contact between the root and the soil may be insufficient to allow transport of water and nutrients between the two (Stirzaker *et al*., 1996).

For the plant roots to elongate in soil, the physical resistance exerted by soil pores against the cross-section of the root must be less than the pressure exerted by the root (Bennie, 1996). The maximal axial pressure exerted by plant roots is between 0.24 and 1.45 MPa, with radial pressures ranging from 0.51 and 0.9 MPa (Misra *et al*., 1986). As pressures increase, plant roots cease to grow through the soil matrix – and show a preference for alternate routes. Increasing bulk density of soils is likely to physically strengthen soils, restricting the development of plant roots. Strizaker *et* al. (1996) showed that an increase in bulk density from 1.12 to 1.78 Mg m^{-3} decreased the root length of barley seedlings in an approximately linear fashion (Figure 5.7). In this experiment, roots grew considerably deeper at bulk densities of 1.5 Mg m^{-3} (180 mm) than at bulk density of 1.77 Mg m^{-1} (80 mm) (Figure 5.8a,b). However, root dry weights were less affected by bulk density than root length, as roots in the higher density soils compensated by growing thicker.

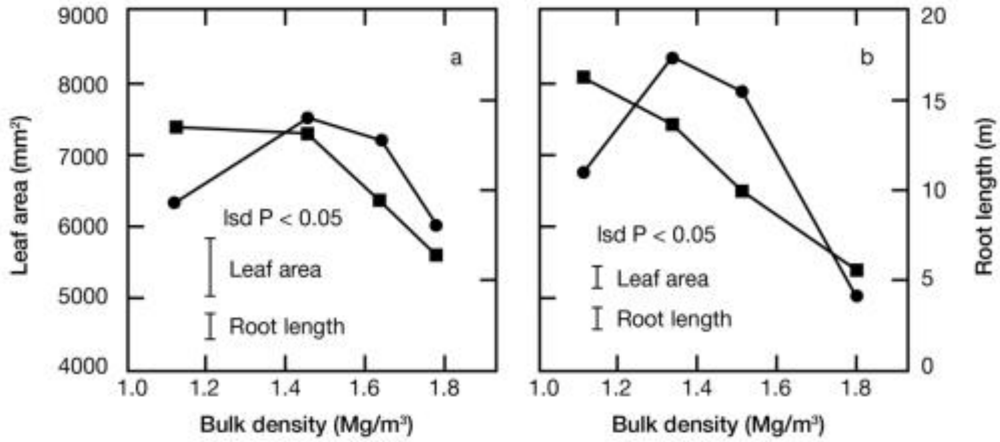

Figure 5.7 Leaf area ● and root length ■ of barley plants as a function of bulk density an (a) continuously wet, (b) alternating wet and dry (from Strizaker *et al., 1996*)

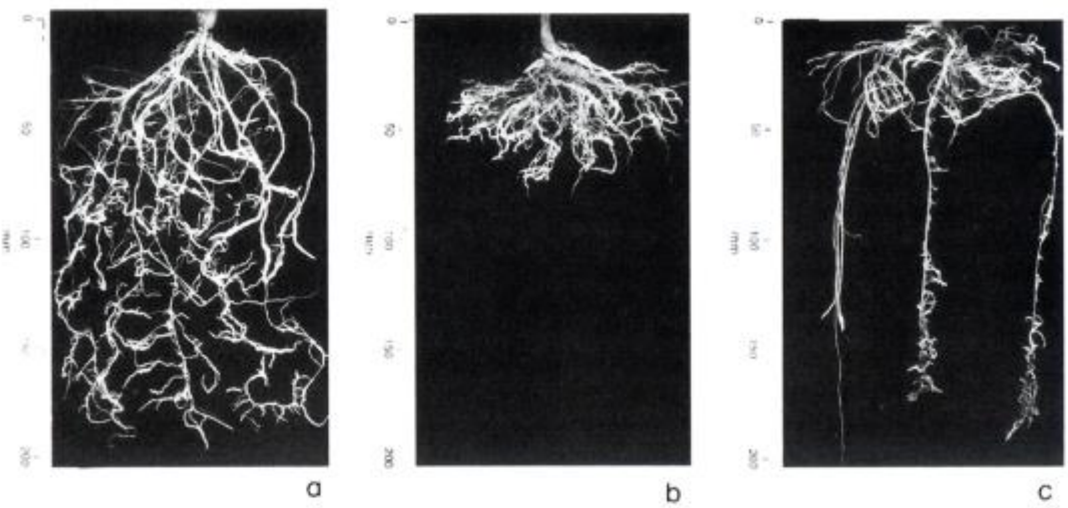

Figure 5.8 Barley roots growing in soil with (a) a bulk density of 1.5 Mg m^{-3}, (b) a bulk density of 1.77 Mg m^{-3}, (c) a bulk density of 1.77 Mg m^{-3} containing 3.2 mm biopores (Strizaker *et al*., 1996)

While compacted layers have high mechanical impedance, they do not have uniform low porosity as they are broken by cracks, faunal channels and biopores. Breaches such as these allow roots an avenue to move through hard layers into deeper layers although the roots remain confined within cracks or biopores (Wang *et al*., 1986). In soils with good structure, that is, have continuous macropores, structural limitations imposed by the soil matrix are largely irrelevant (Lal, 1984). However, while cracks allow roots the opportunity to penetrate otherwise impenetrable layers, roots within the crack tend to clump within the crack rather than explore the greater soil matrix (Figure 5.8a,c). This clumping of roots restricts the ability of the roots to access nutrients and dries the soil in the immediate vicinity of the crack. This decreases the hydraulic conductivity in this zone and restricts water movement from the soil matrix (Passioura, 1991). As well, as discussed previously, the lateral swelling of the root expands the containing pore, compressing adjacent soil pores increasing the bulk density in the zone adjacent to the root. The increased density of the pore wall is likely to impede the ability of lateral roots to penetrate the pore wall (Strizaker *et al.,* 1996). The void between the wall and the root surface of lateral roots in large biopores provides no support for the root as it strikes the biopore wall, causing it to buckle and deflecting the tip of the root, confining it within the pore (Strizaker *et al*., 1996).

The Impact of Management on Soil Organic Matter and Structure

Impact of clearing and cultivation on soil organic matter and the structural form

The clearing of native vegetation and the practice of cultivation have immense effects on the soil. Clearing removes the protecting layer of vegetation and exposes the soil surface to weathering, and reduces the frequency of additions of organic matter to soil.

Cultivation physically disturbs soil, mixing upper and lower layers. The combined effect of clearing and cultivation on soil has little parallel in nature.

Clearing Clearing has only a slight depressive effect on total organic carbon levels in soils when compared with levels in soils under permanent pastures, but the detrimental effect on macroaggregate stability (>2000 µm) is more pronounced (Figure 5.4a). Despite having similar total soil organic carbon levels, English studies suggest that the stability of aggregates after 25 years of pasture had not achieved the level of stability of non-cultivated virgin soils (Low, 1955; 1972). This additional stability of virgin soil aggregates may be influenced by the distribution of organic material and the diverse nature of types of organic material existing in these soils that develop over long time periods.

Cultivation While aggregate stability is related to total organic carbon levels in soil, other fractions of the soil organic pool are also important. Macroaggregates depend for long-term stability on decomposing fragments of plant roots and soil hyphae. However, these fractions are susceptible to tillage. Increasing the frequency of cultivation reduces total root and soil hyphal length and stability of aggregates. In undisturbed soil, total hyphal length was 19 m/g soil compared with 13 m/g soil in cultivated soil; the effect was more drastic in fallowed soil with only 5 m of soil hyphae/g soil remaining (Tisdall and Oades, 1980). Yet while cultivation decreases the total length of soil hyphae when compared to non-cultivated soils, in the absence of fallowing, sufficient hyphae grew in association with the roots of wheat plants to contribute sufficiently to macroaggregate stability (Tisdall and Oades, 1980).

Despite the effect that tillage has on soil organic material and its association with the soil inorganic fraction, cultivation physically alters the shape of soil aggregates. The severity of this effect is influenced by:

- the type of implement used;
- soil conditions that exist at the time that tillage is performed;
- frequency of cultivation; and
- speed of operation.

Aside from the effect of compaction caused by the vehicle traffic, tillage:

- can smear and deform aggregates and cause compaction to occur at the base of the tillage layer (Bowen, 1981);
- breaks the continuity of soil pores within the tilled layer and between the tilled layer and the underlying non-tilled layer, and breaks biopores and faunal channels (Kay, 1990; Hermawan and Cameron, 1993).

Soil moisture content at the time of tillage is the single most important factor controlling the effect of tillage on the soil structural form. The friability of soil is greatest when tillage is carried out at or just below the soil lower plastic limit (Ojeniyi and Dexter, 1979). At moisture contents above the plastic limit, aggregates deform but do not crumble when subject to tillage and hence smearing of the pores occur. In contrast, at moisture contents well beneath the plastic limit, aggregates are less susceptible to deformation and are more prone to shattering.

Mechanical fallowing

Until recently, fallowing was a traditional practice in most cereal growing areas of Australia. Fallows conserve soil moisture and nitrate in the period prior to the sowing of the crop. This occurs as the cultivation associated with traditional fallows kills emergent weeds and breaks the connectivity of soil pores (Ball, 1981). This reduces the loss of soil moisture by soil evaporation and nitrate immobilisation. However the practice of mechanical fallowing affects soil organic carbon levels and soil structure in a number of ways:

- mechanical cultivation enhanced the mineralisation of soil organic material by aerating and mixing the organic litter through the soil. The breaking up of organic litter increases the opportunity and area for microbial attack;
- during fallows, contributions of organic material, such as from leaves or senescing roots, are absent;
- the lack of roots in soil during the plant free period cause a decline in the numbers of surviving symbiotic soil microorganisms such as vesicular arbuscular mycorrhizal fungi and other soil hyphae;
- mechanical manipulation of soil and the lack of plant roots reduces the numbers and activity of soil fauna such as earthworms.

The combination of these factors cause total soil organic carbon levels to decline, affecting soil architecture and stability of the structural form. In addition, the growing of annual crops by conventional practices affect the accumulation of soil organic carbon, as these plants only grow for a portion of the year, only contributing organic litter in the later part of their growth cycle. The aggregative effect of increasing the frequency of annual crops, cultivation and fallowing is the reduction in the accumulation of soil organic carbon and aggregate stability.

Impact of conservation tillage

In recognition of the damage that cultivation causes to soil and the increase in susceptibility of tilled land for erosion, farmers have opted for methods of crop establishment which are less dependent on cultivation. With the adoption of minimum tillage practices herbicides are used to control weeds and specialised sowing equipment and methods are required to handle stubble and to sow into uncultivated soil.

Not all soils are suitable for crop production using zero or minimum tillage as the success of these depend on the soil's capability to withstand the compression caused by zero till machinery and the ability of these soils to self-form a macropore network in the tilled layer. Soils which are prone to compression and have only a low structural resilience are unlikely to perform well in conservation tillage systems as they require tillage to overcome compaction in the sowing layer and to assist in creating a macropore network.

Where conservation tillage replaces conventional tillage in soils that are compatible with conservation tillage, long-term changes may occur in soil properties such as:

- levels of organic carbon and nitrogen;
- porosity and pore size distribution of soils;
- stability of aggregates and associated hydraulic properties.

While total soil organic carbon content is likely to decline under any cropping system when compared with levels under pasture, the decline is more rapid where cultivation is practised, particularly where stubbles are burnt. Table 5.1 shows the change in organic carbon levels from a long-term rotation-cultivation trial (Heenan *et al*., 1995). In all cases, where continuous cropping was practised, an increase in the number of cultivations resulted in a more substantial annual decrease in organic carbon relative to continuous subterranean pasture. The effect was further magnified in all except one case (treatment 4) where stubbles were burnt.

Table 5.1 Slope, t-values for fitted lines of changes in soil carbon levels over time as affected by rotation (adapted from Heenan *et al*., 1995)

Treatment #	Rotation	Stubble Management	Tillage	Slope (kg C/ha/yr)	t-value
1	LW	Mulch	direct drill	-44	-0.52 n.s.
2	LW	Mulch	1 cultivation	-60	-0.66 n.s
3	LW	Mulch	3 cultivations	-179	-2.09 *
4	LW	Burn	direct drill	-115	-1.35 n.s
5	LW	Burn	1 cultivation	-183	-2.05 *
6	LW	Burn	3 cultivations	-250	-2.79 **
7	WW	Burn	3 cultivations	-400	-5.29***
8	WW(+N)	Burn	3 cultivations	-348	-4.50***
9	S(grazed)W	Mulch	3 cultivations	-61	-0.68 n.s
10	S(mulch)W	Mulch	direct drill	-7	-0.08 n.s
11	S(mulch)W	Mulch	3 cultivations	-47	-0.54 n.s

L, lupins; W, wheat; S, subterranean clover

* P<0.05; ** P<0.01; ***P<0.001; n.s not significant

Similar effects of substantially more soil organic carbon under minimum tillage compared to conventional tillage have been reported elsewhere (Smettem *et al*,, 1992; Hermawan and Cameron, 1993; Chan *et al*., 1994).

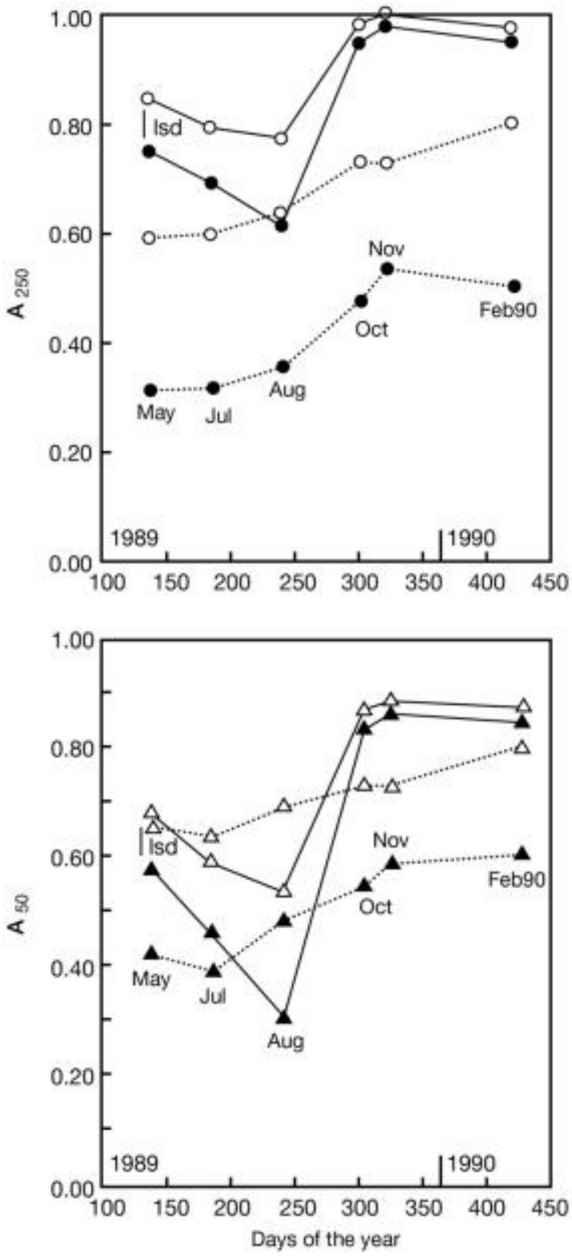

Figure 5.9 Temporal changes in macroaggregate (A_{250}) and microaggregate (A_{50}) stability index under different tillage/stubble treatments: □, DD/SR; □, CC/SB; solid line, -10kPa; dotted line, air dried. A_{250} refers to the fraction of primary particles less than 250 µm remaining as aggregates over 250 µm at the end of the wet sieving and sedimentation treatment. A_{50} refers to the fraction of primary particles less than 50 µm remaining as aggregates over 50 µm at the end of the wet sieving and sedimentation treatment. (from Chan *et al*., 1994)

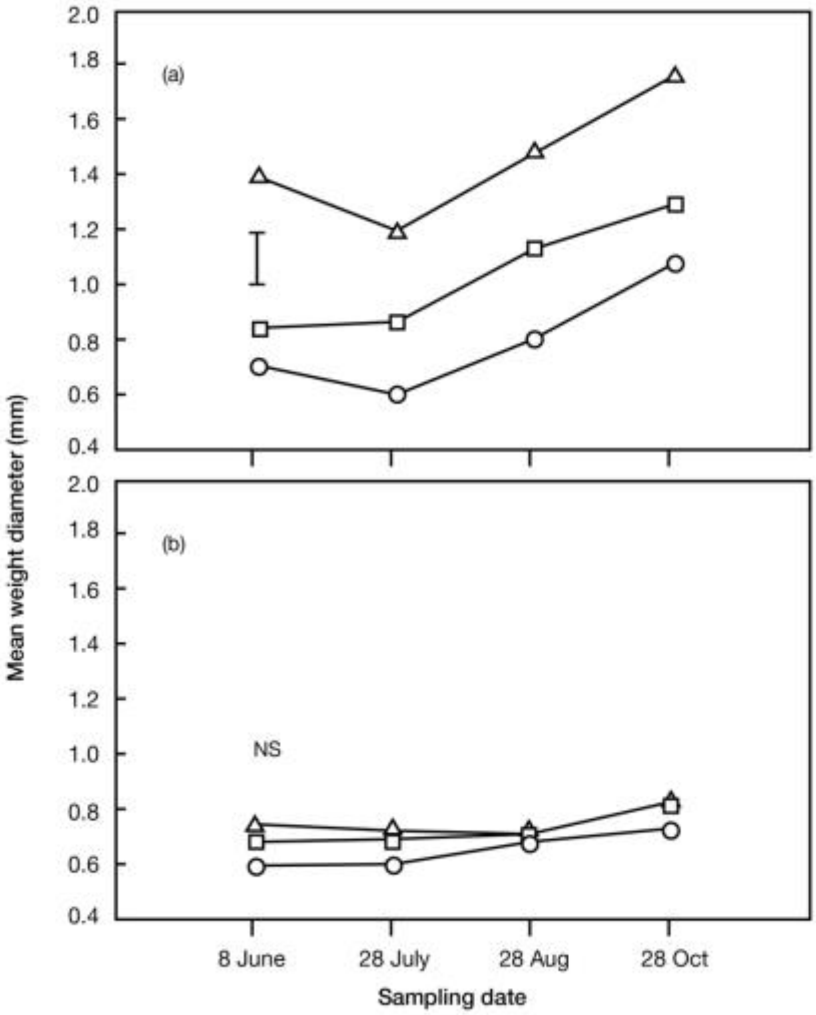

Figure 5.10 Seasonal changes in wet aggregate stability from the surface layer (0-10 cm) of an Oxic Paleustalf with time as affected by tillage treatment: (a) aggregates obtained from the 0-5 cm layer, (b) aggregates from the 5-10 cm layer. The tillage treatments imposed are denoted as: (-▣-) direct drilling, (-□-) reduced cultivation, and (-▣-) conventional cultivation. Vertical space bars in (a) represent LSD values calculated from repeated measure analysis at confidence level of ($P<0.05$). The symbol N.S. in (b) indicates no significant difference at ($P<0.05$) in mean aggregate stability between the three tillage treatments and over time (From Suwardji and Eberbach, 1998)

Improvement in the water stability of aggregates following long-term minimum tillage is well acknowledged (Hermawan and Cameron. 1993; Chan *et al*., 1994; Pagliai *et al*., 1995; Suwardji and Eberbach. 1998). Aggregate stability, however, is not a static soil property but varies through the season. The stability of aggregates in winter-cropped soils decline from late autumn to winter followed by a substantial increase in stability in spring (Figures 5.9, 5.10). This cycling in stability occurs with both macro- and meso-aggregates (Figure 5.9) and is restricted only to aggregates of the surface (0 – 5 cm) layer of soil (Figure 5.10). Chan *et al*. (1994) indicated that this pattern reflected the inverse of soil moisture content at the time of sampling, as they were unable to relate the change in stability with any corresponding change in either total soil organic matter or polysaccharide content. In contrast, Suwardji and Eberbach (1998) suggested that the change in aggregate stability reflected seasonal microbial and plant activity: the autumn - winter decline in stability reflects the decayed state of the pre-season stubble while the flush in spring reflects a flush in plant growth and associated microbial activity. Despite the seasonal variation in aggregate stability, evidence suggests that, at any particular point in time and on otherwise similar and comparable soils, aggregates obtained from conservation tillage are more stable than those obtained from conventionally tilled fields (Figure 5.10).

Long-term conservation tillage affects the geometry of pores and the pore network in soil. While cultivation improves the proportion of irregular shaped macropores in the tillage layer, due to the mechanical shifting of aggregates in relation to each other, long-term conservation tillage improves the continuity of the pore network. More elongated transmission pores (50-500 μm) have been reported in alluvial clays and silt loams where minimum tillage had been practised over several years (Pagliai *et al*., 1995). However this was not observed in pores of similar size (60 – 300 μm) on other soils of similar texture (Hermawan and Cameron, 1993). This contrast may relate to the different mineralogy of the clay fraction (particularly their inheritant swelling-shrinking properties), the type of crops grown and the prevailing climatic conditions. However, improvements in the numbers of storage pores in the size range 0.2 – 60 Δμm are commonly reported on most soils where minimum tillage has been practised over several years (Hermawan and Cameron, 1993; Pagliai *et al*., 1995).

As the soil structural form influences soil hydraulic properties, then fluctuations in aggregate stability as affected by tillage or by season are likely to affect the rate of water infiltration into and conduction through soils as well as gaseous movement through soils. Field observations have revealed a myriad of different results in relation to the impact of tillage on soil hydraulic properties. Hermawan and Cameron (1994) reported that there was little difference in infiltration rate between tilled and minimum tilled soils. This contrasted strongly with other findings which indicated that time for initiation of runoff was greater under minimum tillage (Packer *et al*., 1984). However conditions of surface management may confound these observations. For example, livestock grazing on crop stubble may compact the surface layer of any soil regardless of cultural practice and have a detrimental effect on surface hydraulic characteristics. Other studies show that soils under long term minimum tillage have superior hydraulic characteristics to those in conventionally tilled soils. These improvements are likely to relate to:

- an improved network of macropores and storage pores persisting in the sowing layer; and
- better pore connectivity between the sowing layer and the lower layer (Pagliai *et al*., 1995).

Nonetheless, soil hydraulic properties fluctuate differently during the season as affected by tillage. In some soils, macropores created with cultivation may promote early season saturated hydraulic conductivity (*K*sat) but in the latter part of the season increased plant and soil faunal activity increase *K*sat, particularly in minimal tilled soils

(Suwardji and Eberbach, 1998). This effect was not mimicked in unsaturated hydraulic conductivity (*K*-40) which, regardless of tillage treatment, declined progressively through the season (Figure 5.11). This may be due to the blockage of moisture conducting pores by the development of the plant root system, particularly in spring.

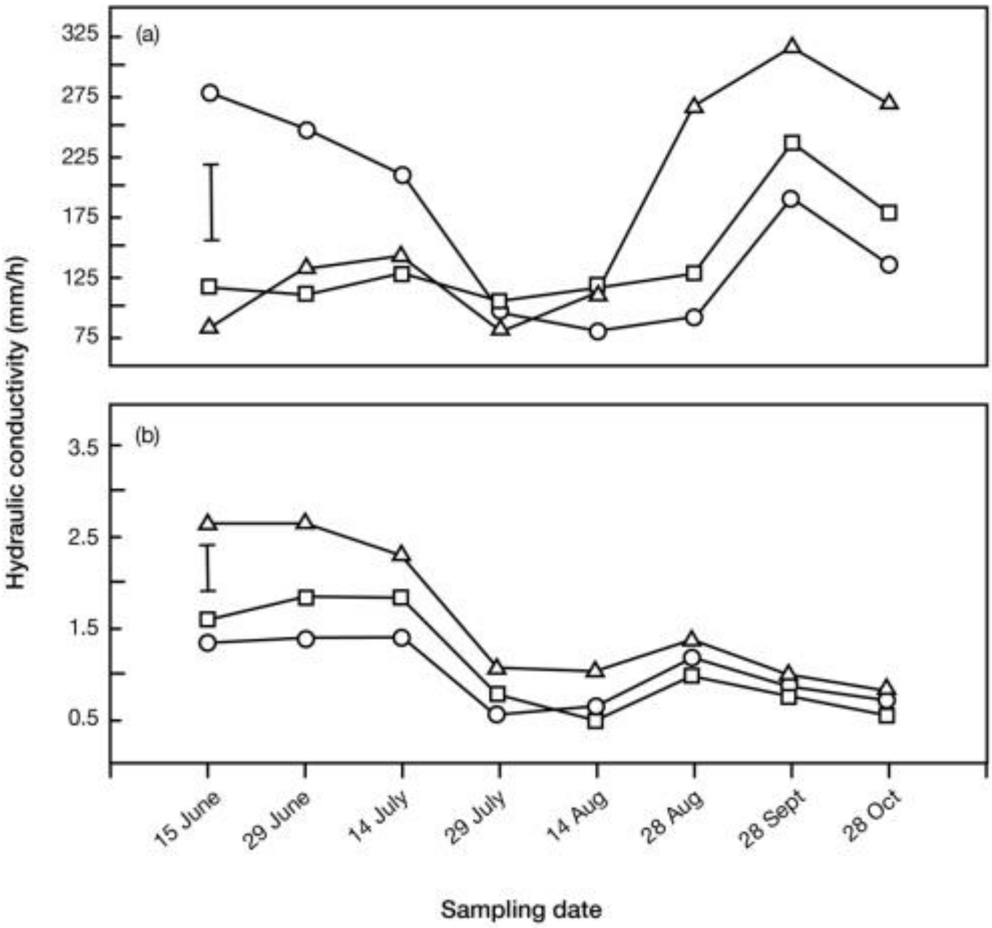

Figure 5.11 Seasonal changes in saturated and unsaturated hydraulic conductivities of the surface layer of an Oxic Paleustalf as a function of time and as affected by tillage treatment. (a) Saturated hydraulic conductivity at a head of 5mm, and (b) unsaturated hydraulic conductivity at a tension of –40 mm. The tillage treatments imposed are denoted as: (-▣-) direct drilling, (-□-) reduced cultivation , and (-▣-) conventional cultivation. Vertical space bars represent LSD values calculated from repeated measure analysis at confidence level of ($P<0.05$). (From Suwardji and Eberbach, 1998)

In soils which are compatible with conservation tillage, the beneficial effect of direct drilling takes some time to occur. In the initial years after adoption of minimum tillage, bulk densities of surface soils (0-15 cm) were higher than those where traditional cultivation techniques were employed (Voorhees and Lindstrom, 1984). However after 3-4 years, this effect was reversed. After 7 years of conservation tillage, the porosity of the 15 –30 cm layer in conservation tilled soils may exceed those where cultivation is employed.

SOIL MANAGEMENT AND THE AVAILABILITY OF WATER FOR CROPS

The cultivation of winter grown crops in Australia is largely restricted to areas with a mean annual rainfall of between 250-800 mm. The climate in these areas ranges from true Mediterranean as experienced in the Western Australian wheat belt to the subtropical climate of northern Australia. In the cereal growing regions of northern Australia, summer rainfall predominates and winter sown cereal crops are grown largely on stored water. This water represents rainfall captured over the previous summer fallow and stored in the subsoil for later use. By contrast, rainfall in southern Australia ranges from being evenly distributed over the year as occurs in central New South Wales to winter dominant as occurs in Victoria and South Australia. Winter sown crops in some districts of southern Australia may benefit from water captured in out-of-season fallows but, in most areas, crops grow on rain that falls during the season (April – October).

As a result of the location of the Australian land mass in the mid-latitudes, and with the prevailing influence of ocean currents, weather patterns in much of Australia's grain growing regions are highly erratic. Rainfall is highly variable and difficult to predict with any great accuracy. As a consequence, crop yields are also quite variable. In the wheat growing areas of southern Australia, winter rainfall generally occurs at a rate that exceeds *potential evapotranspiration* (Ep), while in the period post-anthesis, Ep exceeds rainfall. Hence, while most crops depend on in-season rainfall, they are reliant during the latter part of the season on water stored in the soil during the winter period. This is particularly apparent in the central west of NSW where the yields of spring wheat largely depend on the rainfall for the period 3 weeks either side of anthesis (Seif and Pederson, 1978). It therefore becomes important to manage soils to:

- maximise water infiltration, storage, and the ability of roots to capture stored moisture; and
- minimise the losses of soil water from the system by either *evaporation* or via *deep percolation*.

How then should soils be managed so as to promote efficient and sustainable use of soil water by crops?

Consideration of the *water balance* at a particular location over a particular period is required. The water balance equation most commonly used is as follows:

$$P + U = R + E + T + I + D + L + \Delta S \qquad (1)$$

where *P* is *precipitation*, *U* is *upward capillary water movement* into the root zone, *R* is *surface water movement*, *E* is *evaporation from the soil surface beneath the crop canopy*, *T* is *transpiration* (evaporation from the surface of plant tissue), *I* is *rainfall intercepted by the plant canopy*, *D* is vertical *drainage* of soil water beneath the root zone of the crop, *L* is *lateral sub-surface movement* of soil water, ΔS is the *change in soil water content* between two points over time. In most studies *P* is directly measured, *ΔS* is measured using a neutron moisture meter and *D* can be estimated a number of different ways. Except at unique sites, *U* and *R* are assumed to be zero and *E, T* and *I* are collectively calculated as *evapotranspiration* **(***Et*) by difference, once the other components are estimated. *P, U, I, L* are largely independent of soil management practices and are not considered further here.

Any attempt to maximise the water use efficiency of crops requires the system to be managed agronomically and edaphically so as to maximise *T,* maximise water infiltration and storage in the soil, and to minimise *D* to sustainable levels.

Evapotranspiration in relation to crop yield

The components of evapotranspiration, *E, T* and *I,* are often treated collectively as the separation of the terms, *E* and *T* is problematic. Both are intrinsically a part of *Et* and individual measurement of the terms difficult. Despite these obstacles, it is important that *E* and *T* be independently understood as crops benefit directly from *T* but only indirectly from *E.* In addition, *T* and *E* are not fixed proportions of *Et* but have a reciprocal relationship, with *E* dominating in the early growth stages of the crop and *T* becoming more important as the season progresses. In spite of the difficulty in estimating *E* and *T*, there are numerous techniques by which *E* can be managed and these are considered briefly.

A good relationship exists between crop yield and *Et*, but the timing of availability of soil moisture to meet plant demands is perhaps as important in determining yield as is the gross amount of water used by a crop over a season. The relationship between *Et* and yield is illustrated in Figure 5.12. When water deficit stress is relieved by supplemental irrigation, crop yields improve linearly with *Et*. Figure 5.11 shows also that yield of bread wheat does not commence immediately with *Et* but at a base *Et* value of between 100-200 mm – a loss of water vapour from the system mostly due to *E*.

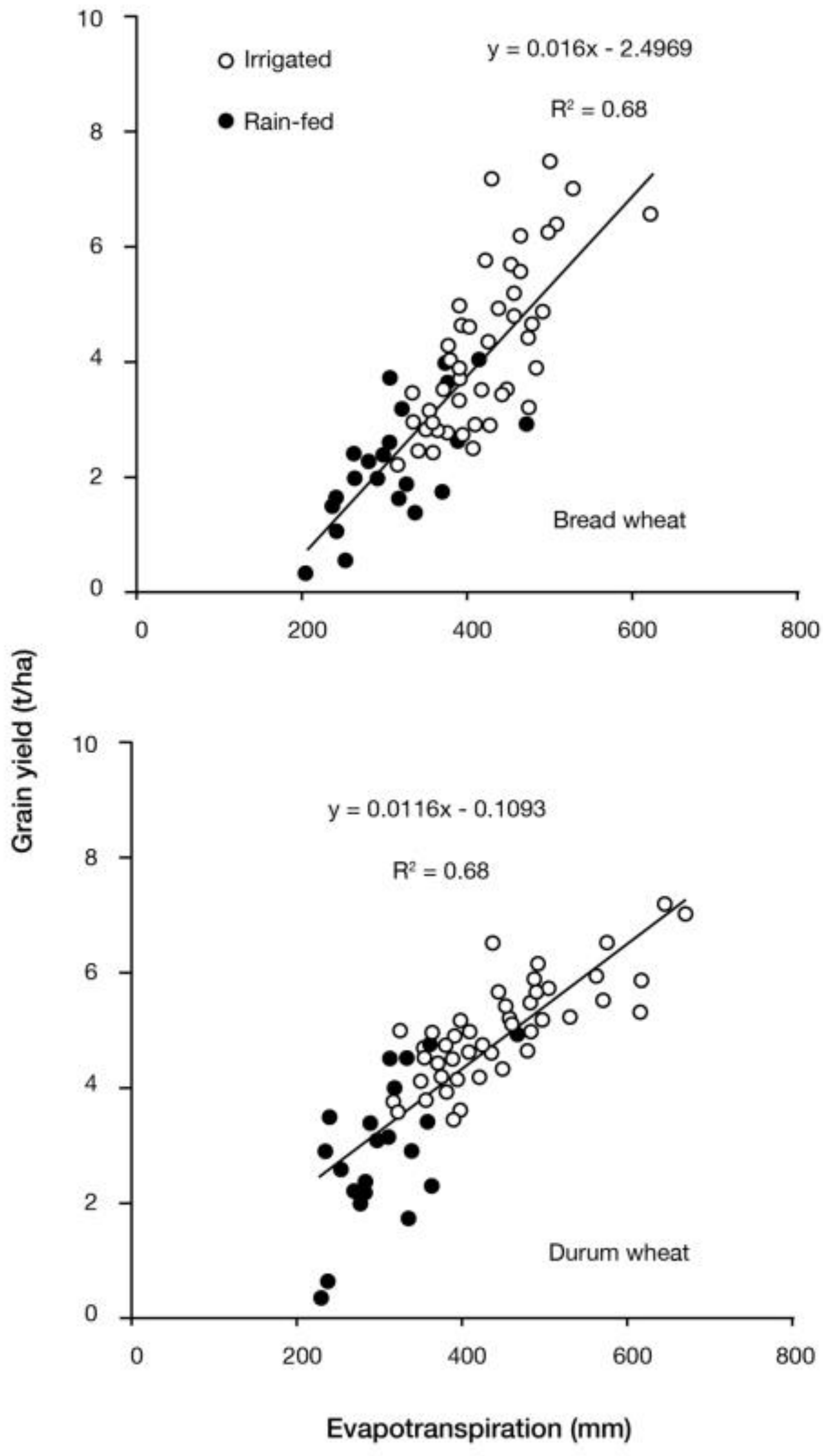

Figure 5.12 Relationship between crop grain yield and evapotranspiration for bread and durum wheat over five consecutive seasons 1991/1992 to 1995/1996 (from Zhang *et al.*, 1998)

Independent of rainfall distribution, French and Schultz (1984) showed that wheat crops in South Australia share a common level of soil evaporation, about 110mm. Other estimates of soil evaporation show that, in reality, this figure is variable, ranging from 70 mm (Siddique *et al*., 1990) to 160 mm (Allen, 1990). This variability in *E* may be due to edaphic and agronomic influences that promote early development of canopy cover or act to reduce the loss of soil water from the soil surface. Other work conducted on soil with high hydraulic conductivity (similar to those used in Zhang and Oweis, 1998) showed cumulative soil evaporation estimates to be close to 200 mm over a crop growth season where 433 mm of rain fell (Eberbach and Pala, unpublished). As crops derive little benefit from soil evaporation, except for a slight moderation in the saturated pressure deficit of the air about the canopy, minimising this loss of soil water increases the amount of water available for the crop for transpiration and generating biomass or yield.

The transpiration process cools plant tissue and provides the plant with a transport medium. It also provides water to act as a solvent for chemical reactions and for photosynthesis. The relationship between transpiration and grain yield is about linear for Australian wheat varieties with the slope of the relationship equivalent to about 20 kg grain/ha/mm water transpired (Passioura, 1976). This figure provides the genetic upper limit to transpiration efficiency for current wheat varieties. Analysis of yields taken from farms across southern Australia indicates that this limit is rarely achieved (Hamblin *et al*., 1987; Cornish and Murray, 1989). While there are many agronomic and edaphic reasons for this, climate and the timing of availability of soil moisture in relation to critical plant periods are likely to be significant. Soil management can provide opportunities to moderate *E* and improve the amount of water accessible by roots for *T* . Therefore any technique that enhances the development of the plant canopy and decreases the transmission of radiation and air movement at the soil surface will moderate evaporation. Examples include:

- improved soil chemical fertility which promotes early vegetative development of the plant canopy and promotes shading;
- adoption of stubble retention practices that create a mulch layer. Mulches
  - maintain warmer soil temperature in late autumn to enhance germination and plant development;
  - act to reduce wind speed and intercept radiation; and
  - physically impede water vapour loss.

However, stubble retention practices may:

- hinder sowing operations;
- slow the early development of crops; and
- harbour plant diseases.

All of these affect crop production and performance in other ways.

Whereas managing soils to restrain losses of soil water by soil evaporation indirectly contributes water for transpiration, soils can be managed to augment directly soil water for transpiration. Techniques which may improve the amount of soil water available for transpiration, and the ability of plant roots to capture this water, include:

- increasing the proportion and stability of macropores in the surface sowing layer to aid infiltration, redistribution of soil water in the unsaturated phase and to aid drainage;
- improving the continuity of soil pores between the sowing layer and subsurface layer thereby aiding water movement deep into the soil away from where it may evaporate and to a position in the soil profile where it may be accessed later in the season by plant roots;
- improving the number of water storage pores in soil so as to increase the soil's volumetric store of soil moisture;
- improving the network of biopores and faunal channels to ease the passage of roots through the soil thereby enabling roots to grow deeper, explore soil more extensively and capture more soil water.

These improvements rely on altering management in two ways. Stubble retention reduces the rate of movement of water across the soil surface. This protects the soil surface, aids soil structural stability and reduces erosion, and aids infiltration (Mason and Fischer, 1986; Norwood, 1994). Adoption of reduced tillage or direct drilling improves structural integrity and stability, enhances pore continuity and the creation of networks of biopores. It also improves the number of water holding pores (Pagliai *et al*., 1995), encouraging water to move deeper in the profile (Norwood, 1994)

Fallows have traditionally been used by farmers to accumulate water in the root zone prior to establishment of a crop. Mechanical fallows have declined in popularity, being replaced by chemical (no-till) fallows in areas that benefit from this practice. No-till fallows have been shown to store more soil water at depth than mechanical fallows (Norwood, 1994) and the improved moisture status has translated well into higher grain yields in both wheat (Smika *et al*., 1990) and sorghum (Norwood, 1994). However this experience is not universal as a comparison of fallowing methods in the Lockhart region of New South Wales produced similar outcomes for water storage and crop yield following chemical and mechanical fallows (Mason and Fischer, 1989)

Drainage

While management techniques exist to enhance infiltration and hence the amount of water stored at depth, it is important that crops are managed to encourage deep usage of soil moisture and prevent its movement from beyond the root zone. This theme is particularly topical because this leakage of water below the root zone contributes to groundwater recharge, elevating the height of these subterranean water bodies and mobilising salts into the root zone. Soil management techniques to reduce this component of loss include:

- improving soil management options to promote the early vigour of plants so that they establish deep and effective root systems early; and
- managing soils to promote the network of pores to ease the extent and depth to which roots can grow to capture more soil water.

SOIL pH

For the majority of crops, maximum yield is achieved in the soil pH range of 5.5-6.5. Yields decrease below and above this range. However, there are exceptions. For example, lupins and triticale perform well in more acidic soils whilst medics such as lucerne prefer alkaline soils. In Australia, 13.6 million hectares of agricultural land have a pH_{Ca} of less than 4.8 (Anon., 2001). The problem of low soil pH occurs in regions of rainfall in excess of 500 mm per annum and in irrigated areas. The problem of high pH is common in lower rainfall environments with calcareous sands and cracking clays as well as with many sodic non-saline soils.

Table 5.2 Estimated areas of agricultural land in Australia having specified pH_{Ca} ranges (Anon., 2001)

pH range	Area	
	millions of hectares	percentage of agricultural land
<4.3	0.3	0.3
4.3-4.8	13.4	15.0
4.8-5.5	37.1	41.0
5.5-7.0	24.0	26.9
7.0-8.5	14.4	16.2
>8.5	0.001	0.001

In Australia, soil pH is measured in a 1:5 soil:solution extract ratio using either distilled water or 0.01 M $CaCl_2$. The relationship between the two extractants tends to be linear, for example:

$$pH_{Ca} = 1.25pH_w - 2.10 \qquad R^2 = 0.91$$

where pH_{Ca} is the pH in 0.01 M $CaCl_2$ and pH_w is the pH in water (Slattery *et al*., 1995).

In acidic soils, the pH in $CaCl_2$ is normally lower than in water by approximately 0.5-1.0 pH units. (Note: The pH in $CaCl_2$ can exceed the pH in water if the soil has a net positive charge).

Causes of pH Extremes

Soil pH changes only if there are changes in the H^+ or OH^- concentrations. Within soils there are numerous chemical and biological processes that release or consume these ions and any pH change will be the result of the net change in H^+ concentration. The major processes influencing soil pH are commonly associated with the C and N cycles. Conyers *et al.* (1995) described most of the major processes in detail and these are summarised below.

Acidifying Reactions

Nitrification The common nitrification process involves the oxidation of NH_4^+ from fertiliser sources or mineralisation of organic matter, to NO_3^- by chemoautotrophic bacteria. Nitrification occurs in two stages that can be summarised as:

$$NH_4^+ + 2O_2 \rightarrow NO_3^- + H_2O + 2H^+$$

Humification As organic matter is humified there is an increase in the carboxyl functional groups that can dissociate releasing protons. The process assumes importance during the accumulation of organic matter under pastures.

$$R - CH_2OH + O_2 \rightarrow R - COO^- + H_2O + H^+$$

Oxidation of S and sulphides When elemental S is applied as a fertiliser, the S is oxidised by *Thiobacillus* spp. releasing protons.

$$2S + 2H_2O + 3O_2 \rightarrow 2SO_4^{2-} + 4H^+$$

This also occurs when waterlogged soils high in sulphides are drained giving rise to acid sulphate problems. Acid sulphate soils are more frequent in coastal areas being drained and developed for sugar cane production, dairying or urban expansion.

Oxidation of Fe and Mn When waterlogged soils drain any reduced Mn or Fe will be oxidised releasing protons.

$$Mn^{2+} + 1/2O_2 + H_2O \rightarrow MnO_2 + 2H^+$$

Alkaline Reactions

Accumulation of bicarbonate/carbonate Accumulation of bicarbonate/carbonate ions or their application in liming materials results in consumption of H^+ and elevation in soil pH. If $CaCO_3$ is the dominant form the pH will not exceed 8.3 as above this pH the compound becomes insoluble. Higher pH values are commonly associated with Na_2CO_3, as occurs in the alkaline group of the sodic non-saline soils. The reaction of agricultural lime may be represented as:

$$CaCO_3 + 2H^+ \rightarrow Ca^{2+} + H_2O + CO_2$$

Mineralisation of N and urea hydrolysis During decomposition of organic matter by mainly heterotrophic microorganisms, N is mineralised to NH_4^+. In this process, amine groups, for example, are converted to NH_3 which, at pH values below 8, is hydrolysed to ammonium.

$$R_1 - CH(NH_2) - R + 0.5O_2 \rightarrow R_1 - CO - R + NH_3$$

$$NH_3 + H_2O \rightarrow NH_4^+ + OH^-$$

where R and R_1 are organic groups.

Urea is rapidly hydrolysed to NH_4^+, by the enzyme, urease.

$$(NH_2)_2 CO + H_2O \rightarrow (NH_4)_2 CO_3$$

Denitrification As aeration is reduced following waterlogging, NO_3^- is used as the electron acceptor in place of O_2 and in the process is reduced to nitrogen gas.

$$2NO_3^- + 12H^+ + 2e^- \rightarrow N_2 + 6H_2O$$

Association then Oxidation of Organic Anions Return of crop residues or decreases in soil organic matter concentrations during cropping after a pasture phase normally result in an increase in soil pH following association and subsequent oxidation by microorganisms.

$$R - COO^- + H^+ \rightarrow \quad R - COOH$$
$$R - COOH + O_2 \rightarrow \quad CO_2 + H_2O$$

The extent of any pH change induced by addition of organic anions is dependent on the pK of the organic acid and soil pH (Ritchie and Dolling, 1985) and the rate of return (Paul *et al*., 2001b).

Reduction of Fe and Mn In waterlogged and severely compacted soils, gas exchange between the soil and the atmosphere is restricted. The oxygen concentration decreases and Fe and Mn are used by microorganisms as electron acceptors resulting in consumption of H^+.

$$MnO_2 + 4H^+ + 2e^- \rightarrow Mn^{2+} + 2H_2O$$

Balance in the Uptake of Anions and Cations by Plants The processes described above do not include any effects induced as a result of nutrient uptake by growing crops. The following discussion provides a simplified example of the effect of plant uptake of the nutrients. Only nutrients taken up in high amounts (Table 5.3) are included as these have the dominant effect. Also included in Table 5.3 is the concentration of charge accumulated as a result of the uptake of each ion. Nitrogen is not included as it can be taken up as an anion (NO_3^-), a cation (NH_4^+) or uncharged N_2. Assuming an N concentration of 2%, this is equivalent to an uptake of 143 m moles of charge per 100 g. .

Table 5.3 Average dry matter concentration of cations and anions taken up in high amounts and calculated uptake of charge

Element	*Concentration in dry matter (%)*	Charge uptake (m moles /100 g)
K	2.0	51
Ca	1.0	50
Mg	0.2	17
Na	0.1	4
	Positive charge uptake as cations	122
P	0.3	10
S	0.3	19
Cl	0.15	4
	Negative charge uptake as anions	33

The following calculations are provided to demonstrate the effect of N source on pH changes. The magnitude of any effects will vary depending on the actual nutrient content of the crop, which can vary with the form of N taken up. For the purpose of this example it is assumed that the form of N does not influence the balance of nutrients taken up by the plant.

During nutrient uptake there is normally an imbalance of charge influx depending on the balance of cation to anion uptake. This imbalance may be addressed by an outflow of protons or alkali generated within the plant from dissociation of organic acids such as malic acid. This dissociation may be written as:

$$\mathrm{R-COOH} \Leftrightarrow \mathrm{R-COO^{-} + H^{+}}$$

If all N is taken up as the anion, nitrate:

- *an alkaline residue remains in the soil.* The negative and positive charge taken in becomes 176 (143 + 33) and 122 m moles/100 g respectively. The excess uptake of charge as anions over cations is balanced by the excretion of OH^- (54 m moles/100 g) or other alkali such as the organic anion or bicarbonate;
- *the cytoplasm in the plant tissue has an excess of alkali.* The reduction of nitrate to ammonium following uptake consumes 143 m moles/100 g H^+ and leaving 143 m moles/100 g of OH^- (or equivalent as organic anion). There were 54 m moles/100 g of H^+ left from the dissociation of organic acid and the excretion of OH^- into the soil so there is a net alkali of 89 (143-54) m moles /100 g left in the plant.

If all N is taken up as the cation, ammonium:

- *an acidic residue remains in the soil.* The negative and positive charge taken in becomes 33 and 265 (143 + 122) m moles/100 g respectively. The excess uptake of charge as cations over anions is balanced by the excretion of 232 (265-33) m moles H^+/100 g generated by organic acid dissociation;
- *the cytoplasm in the plant tissue has an excess of alkali.* The conversion of ammonium to amino compounds following uptake releases 143 m moles/100 g of

H^+. There were 232 m moles/100 g of alkali as organic anion left from the excretion of H^+ into the soil so there is a net alkali of 89 (232-143) m moles/100 g left in the plant.

The consequences of plant uptake of ions from the soil are:

- removal of plant products results in removal of alkali from the soil usually in the form of organic anions such as malate. For example, export of 3 t lucerne hay/ha, 1 t wheat/ha and 6 kg wool/ha remove the equivalent 350, 10 and 1 kg lime/ha respectively;
- if the crop residues are returned to the soil surface and the soil is not mixed by cultivation,
    - the surface of the soil will have a higher pH following association and oxidation of the residues; and
    - the layers below will be acidic from the excess of cation over anion uptake.

The application of urea fertiliser produces the following combination of processes in the nitrogen cycle to affect soil pH (Box 6.1):

BOX 6.1 The influence of soil N transformations and plant uptake on soil pH following the application of urea fertiliser

The following is an example of the combination of these processes in the N cycle that occurs following the application of urea fertiliser.

- Hydrolysis of urea fertiliser to NH_4^+ results in the release of 1 mole of OH^- per mole of N applied.
- Nitrification of NH_4^+ to NO_3^- results in the release of 2 moles of H^+ per mole of N.
- Plant uptake of nitrate results in the release of alkali.

There is negligible acidification if all urea-N proceeds through this sequence to the plant.

The soil will acidify if the N was leached from the surface soil after nitrate production.

Field studies have measured each of these processes under various management systems and attempted to estimate the relative importance of each process to pH change. Poss *et al.* (1995) concluded that acidification was negligible under wheat grown in the Riverina provided that stubble was returned to the soil (to enable the alkaline effect of organic anion association and oxidation), nitrate was not leached beyond the root zone (a potential acidifying effect) and urea fertiliser was recovered by the crop. In contrast on sandy soils in Western Australia, Dolling *et al.* (1994) found acidification to a depth of 60 cm under wheat and rotations. This was due mainly to leaching of nitrate from mineralisation and fertiliser, removal of products and, where a pasture phase was included, a build-up of organic matter. The sandy soils would have had a lower buffer capacity and be more prone to leaching than the clay soils of the Riverina. Paul *et al.* (2001) showed that the surface 2 cm of soil became more alkaline than the soil below this depth under wheat and subterranean clover because of the return of residues to the surface. Under a fallow treatment, the surface soil was more acidic because any organic N, that was mineralised and nitrified, leached and there was no alkali from nitrate uptake or from return of plant residues. Under grazed pasture, Ridley *et al.* (1990) found acidification rates were higher under limed and P fertilised treatments than unamended soils. In the limed or fertilised pasture, there was more (1) product removal (wool, hay, meat), (2) product transfer (dung and urine), (3) organic matter accumulation and (4) leaching loss of nitrate produced from mineralisation and nitrification.

Poor crop growth at pH extremes

Poor crop performance is the result of a complex of factors, rarely only one. The following discussion highlights the factors involved.

Hydrogen ion toxicity While pH measures the H^+ concentration in the soil, H^+ toxicity rarely restricts root growth. For example, in soils, lucerne growth is restricted below pH 5.5. However, Andrew (1976) showed that in solution culture where only the essential plant nutrients were added at the concentrations that were optimal for plant growth, lucerne grew well down to pH 4.5. At pH values less than 4, protons may damage cells within the roots.

Aluminium toxicity Plants do not require Al for growth but accumulation of Al is toxic to the growing plant. Aluminium toxicity reduces branching of lateral roots and induces P deficiency (Helyar, 1978).

Aluminium toxicity is a major problem in acidic soils as Al solubility increases as soil pH decreases. Data from 4 soils in northern Victoria shows that above pH_{Ca} 5.0 little extractable Al was present in any soil (Figure 5.13). Below this pH, the concentration increased rapidly. While Al toxicity is normally associated with acidic conditions, it has been shown that where a soil high in soluble Al is overlimed to a pH greater than 7, increased Al uptake and reduced growth of corn has been observed (Farina *et al*., 1980). This may be due to the increased solubility of Al as pH increases above 7 (Black, 1968).

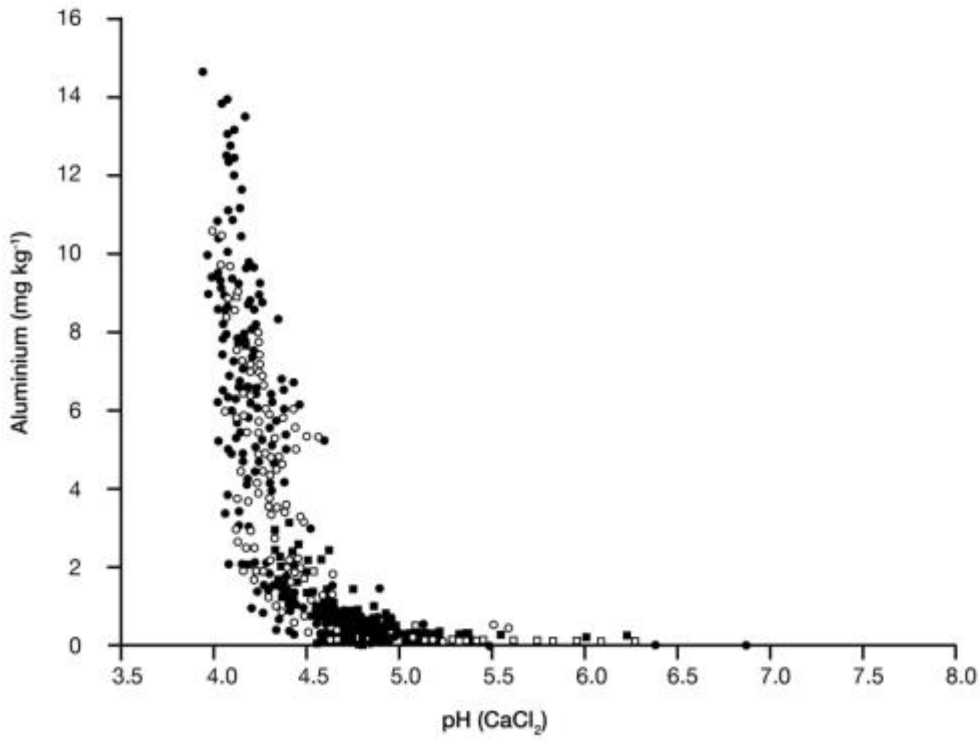

Figure 5.13 The relationship between pH_{Ca} and 0.01 M $CaCl_2$ extractable Al (from Slattery *et al*., 1994)

Toxicity or availability of micronutrient cations Manganese, Fe, Zn and Cu are soluble at low pH but become less soluble as pH increases (Lindsay, 1972). The trends in the availability of Cu and pH are not always apparent (Loneragan, 1975) as Cu concentrations in solution may be maintained through the formation of copper-organic matter complexes.

In soil with low pH and

- high total concentrations of the micronutrient, toxic concentrations develop. Slattery *et al*. (1994) showed that extractable Mn concentration in soil was low above pH 6 but increased rapidly below pH 5. Mn toxicity is one of the major limitations in acidic soils (Fenton and Helyar, 2000);
- low total concentrations of the micronutrient, deficiencies develop. The high solubility results in leaching loss. In sandy acidic soils in southern and Western Australia, deficiencies of Cu and Zn are common (Donald and Prescott, 1975).

In soil with high pH,

- the low solubility of these nutrients leads to deficiency. On Vertosols in northern Australia and Calcarosols in Victoria and South Australia deficiencies of Zn, Mn and Cu are common (Donald and Prescott, 1975). Iron deficiency occurs on highly calcareous Vertosols especially in irrigated areas in the Murray Valley in citrus orchards where high pH may occur in combination with high bicarbonate concentrations.

Availability of molybdenum The availability of Mo decreases as soil pH decreases so that deficiency is common in soils with pH below 5.5 (Fenton and Helyar, 2000). In some soils lime application may overcome molybdenum deficiency but in other soils the total molybdenum concentration is so low, lime cannot improve availability adequately.

Availability of phosphorus The availability of P is strongly controlled through adsorption reactions to silicate clay and oxide surfaces. Generally adsorption on Fe and Al oxide surfaces decreases as soil pH increases. In alkaline soils solubility of P is controlled by reaction with calcium compounds and their solubility decreases as pH increases. The net effect is that in many soils the maximum availability of P occurs in the pH range 5.5-7.0.

The effect of increasing pH in acidic soils through the use of lime has been found to increase, have no effect, or decrease the availability of P (Haynes, 1984). Liming may increase P uptake by crops by reducing Al toxicity. However if extractable aluminium concentrations are high, the freshly precipitated aluminium may adsorb phosphate reducing its availability in the neutral pH range (Haynes, 1984).

Availability of calcium, magnesium or potassium In acidic soils, deficiencies of the cations may occur. At low pH the cation exchange capacity (CEC) of soils is reduced and the higher concentrations of Al on the remaining CEC increases the likelihood of leaching losses of the cations. Deficiencies of Ca have been reported on the east coast of NSW (McLaughlin, 1980) and Queensland (Bruce *et al.*, 1988).

Biological effects on soil pH can have significant effects on biological processes in soils:

- *Mineralisation of nitrogen and nitrification*

Microbial activity is reduced at high and low pH. Liming of acidic soils increases mineralisation of N and subsequent N uptake by grass in soils where growth was not limited by Al^{3+} toxicity (Edmeades *et al.*, 1981). Decreases in N mineralisation (Purnomo *et al*., 2000) and nitrification (Young *et al.,* 1995) through the surface 10 cm of soil of cropping soils have been related to the decrease in pH with depth.

- *Symbiotic nitrogen fixation*

Soil pH extremes influence the survival of *Rhizobium* spp. in soil, their ability to infect roots and the resultant N fixation by the nodules formed on the legume roots. The nature of the effect depends on the N fixing species. The faster growing acid producing *Rhizobium* spp. commonly perform poorly in acidic soils while the slower growing alkali producing *Bradyrhizobium* spp. are more tolerant of low pH. The reverse is true at high pH.

- *Root disease*

Altering pH may increase or decrease disease problems. Low pH can inhibit root diseases such as take-all in wheat (*Gaeumannomyces graminis*) and scab on potatoes (*Streptomyces scabies*). The failure of wheat to respond to liming on acidic soils has been attributed to an increase in the incidence of take-all (Murray *et al.*, 1987). In contrast, liming reduces club root on brassica crops as low pH favours the development of the fungus *Plasmodiophora brassicae*.

Management of pH extremes

Management of low and high pH soils can be achieved by altering the pH with the appropriate amendment. However, there are opportunities to reduce the inputs of these amendments once the principal processes inducing pH change have been identified and strategies developed to control the processes. It is also possible to treat the specific nutrient deficiencies induced by high and low pH or utilise crops that tolerate toxicities.

Alter soil pH

The common materials used to increase soil pH are listed in Table 5.4. Soils may be acidified with ammonium sulphate, aluminium sulphate and elemental sulphur following nitrification, Al hydrolysis and S oxidation respectively.

Table 5.4 Common liming materials and their properties

Common Name	Formula	Neutralising Value #	Comments on use
Agricultural lime	$CaCO_3$	85-100	Low solubility, variable composition
Dolomite	$CaCO_3$. $MgCO_3$	95-108	Low solubility, variable composition, Mg source
Burnt lime	CaO	150-175	More soluble, caustic to handle, fine particle size
Slaked lime	$Ca(OH)_2$	120-135	High solubility, caustic to handle, fine particle size

The neutralising value of a liming material is a measure of its ability to change pH. Pure $CaCO_3$ is used as a reference material and is given a neutralising value of 100.

The common liming material, agricultural lime, has a low solubility. To improve response the following are necessary:

- *finely grind lime* to less than 75 μm (Conyers *et al.*, 1996); and
- *incorporate lime into the soil* as movement through the soil is slow, that is, normally less than 1 cm/year.

Lime rates may be estimated from a combination of:

- the current soil pH as determined by soil testing;
- the buffer capacity of the soil which increases with clay and organic matter content; and
- the desired pH which depends on the crop species being grown.

Alternatively, rates may be selected to match the acid inputs. Cregan and Scott (1998) provided a detailed summary of lime requirements (kg/ha/year) to overcome acidification for various farming systems. Some examples are:
- grazed pasture (NE Victoria), 39
- continuous wheat (NE Victoria), 230
- annual crop/pasture (Wagga Wagga), 46-95

- grass pasture for hay (Queensland), 310
- sugarcane (Queensland), 170.

Excessive rates of lime reduce yield (Scott *et al.*, 1997) by inducing the problems associated with high pH.

Lime responses may be limited by subsoil pH especially when repeat applications of lime are required. Scott *et al.* (1997) showed that response to lime by an acid sensitive barley was minimal where the pH of soil from 10-20 cm was 4.5 (Figure 5.14). Liming the surface 10 cm to pH 6 only increased yield to 1 t/ha. Where the pH of the 10-20 cm soil was 5.5, liming the surface to pH 6 achieved a yield of 2.5 t/ha.

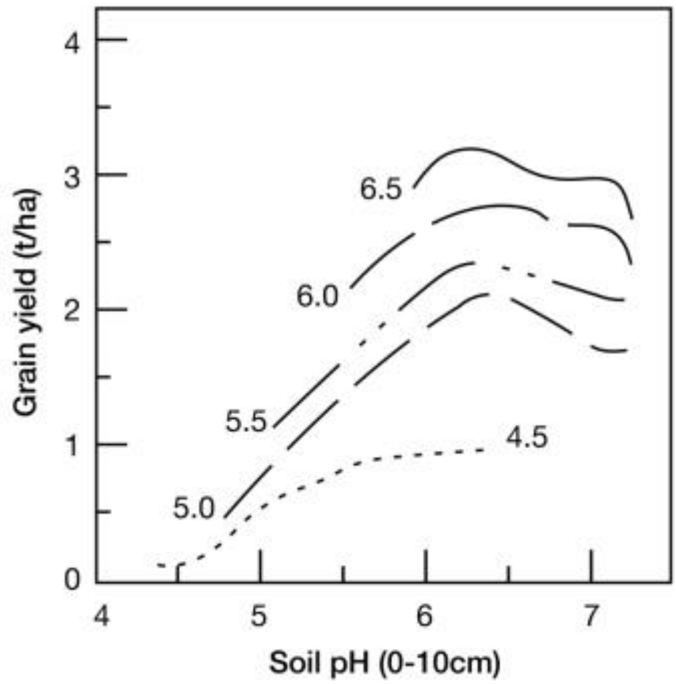

Figure 5.14 Lime response to acid sensitive barley to varying soil pH in the 0 to 10 cm depth. Each line represents the response at different soil pH in the 10 to 20 cm depth (after Scott *et al.*, 1997)

Lime pelleting of legume seed following inoculation overcomes problems associated with the survival of the *Rhizobium* spp. in the soil (Loneragan *et al.*, 1955). Lime pelleting with *Bradyrhizobium* spp. must be avoided, as these bacteria do not tolerate the high pH.

Nitrate fertilisers increases soil pH. Henzell (1971) found that, following the application of 448 kgN/ha for 6 years, the pH changed from 5.13 in the control to 4.15, 4.82, 5.25 and 6.10 with sulphate of ammonia, urea, ammonium nitrate, and sodium nitrate respectively. While sodium nitrate increased the pH by approximately 1 pH unit, the technique would be uneconomic.

Gypsum is rarely effective for amelioration of acidity**.** In soils with a high CEC, gypsum addition increases the effects of acidity (Black and Cameron, 1983). However on soils with a low CEC or an anion exchange capacity, gypsum has achieved improved yields in soils with a high Al^{3+} concentration (Farina and Channon, 1988). The weathered and acidic Ferrosols in northern Australia are probably the only cropping soils with these properties.

Minimise processes contributing to pH change

Techniques such as reducing the extent of NO_3^- leaching, the use of less acidifying N fertilisers (Table 5.5), and the return of plant product containing alkaline products (hay and silage) have been suggested to reduce the rate of acidification (Cregan *et al*., 1989).

In southern Australia, management of NO_3^- leaching has been achieved through the early sowing of crops (Strong, 1992) and the use of deep rooted perennial grasses to recover leached nitrate. However, this was only effective in north-eastern Victoria in drier than average years (Ridley *et al*., 1999). These strategies may reduce the overall acidification rate in the root zone but they may not prevent redistribution of acidity within the root zone. The surface soil may become acidic through the process of mineralisation followed by nitrification and NO_3^- leaching. The subsoil would become more alkaline as a result of NO_3^- uptake. Black (1992) showed that in soil receiving urea application the shallow subsurface (2-6 cm) became acidic while below 8 cm the soil became alkaline.

Table 5.5 Soil acidification rates expected from various forms of nitrogen fertilisers (Cregan and Helyar, 1986).

Fertiliser and Acidification Class	$CaC0_3$ required (kg lime/kg N) to balance acidification where leaching removes the following percentage of applied N	
	0	100
Most acidifying – ammonium fertilisers:		
Sulphate of ammonia (ammonium sulphate), MAP (monoammonium phosphate)	3.7	7.1
Medium acidification:		
DAP (diammonium phosphate)	1.8	5.3
Low acidification:		
Urea		
Ammonium nitrate	0	3.6

Aqua ammonia		
Anhydrous ammonia		
Alkaline fertilisers:		
Sodium and calcium nitrate	-3.6	0

Fertiliser use

Deficiencies of P, Mo, Ca, Mg, Cu and Zn occur in soils with low pH and Mn, Zn, Cu and Fe at high pH. Rather than applying tonnes of amendment (e.g. lime) per hectare to change pH, application of low rates of fertiliser may improve crop growth. For example, application of a few hundred grams of Mo fertiliser has been shown to increase growth of legumes in acidic soil as much as several tonnes of lime.

At high pH, application rates have to be several fold higher than that in neutral soils to achieve the same crop yield. In highly alkaline soils, foliar applications of manganese and iron have been used to avoid the difficulties of ensuring the nutrient is available in the soil.

Use of tolerant crops and varieties

Within and between crop species there is tolerance to Al and Mn toxicities (Table 5.6). Consequently, rather than changing soil pH, species tolerant of the specific toxicity may be included in the farming system.

Table 5.6 Degree of tolerance of common crop and pasture plants to toxicities of Mn and Al (Anon., 1999)

Sensitivity	Toxic element	
	Mn	Al
Highly sensitive	lucerne, pigeon pea, barrel medic, burr medic	lucerne, barley, medics, canola
Sensitive	white clover, strawberry clover, chickpea, canola	red clover, phalaris, sub clover, wheat
Tolerant	sub clover, cotton, cowpea, soybean, wheat (Matong, Vulcan, Dollarbird), Barley (Yerong, Lara, Scooner), triticale (Empat, Muir, Tahara)	ryegrass, tall fescue, cocksfoot, rose clover, fodder rape
Highly tolerant	rice, sugar cane, tobacco, sunflower, oats, most pasture grasses	lupins, oats, triticale, cereal rye, Maku lotus

The disadvantages of using tolerant species are:

- the range of crops tolerant of low pH is restricted so that fewer options are available for rotations; and
- acidification continues. Removal of plant products at harvest removes alkali, that is, an acidic residue is left in the soil.

CONCLUSIONS

Climate and the fertility of soils have a major primary influence on crop yield potential. In this chapter, the influence of soil structure on the plant root environment, its impact on water availability to plants, and the causes and management of pH in the soil have been identified and discussed.

The development and maintenance of the basic soil structural unit is clearly affected by how soils are managed. Methods of crop management that minimally disturb the soil and retain stubble favour the development of a stable-continuous pore network. Such a pore network improves infiltration and drainage, water retention as well as aeration.

The optimum pH for growth varies between species. The adverse effects of high and low pH are caused by a complex array of factors including nutrient deficiencies and toxicities, Al toxicity and unfavourable biological conditions. Understanding and identifying each of these limitations provides management opportunities in acidic or alkaline soils.

PRINCIPLES

- The movement of plant roots, soil biota, and soil solids are the primary contributory processes to aggregate and pore development.
- Fresh and decaying plant roots and fungal hyphae are major contributors to aggregate binding but are vulnerable to soil management.
- Layers of high soil bulk density, occurring either due to aggregate slumping or compaction, create conditions of high mechanical impedance to root movement and influence the entry and movement of water and air.
- Crop establishment practices such as direct drilling and stubble retention minimise loss of the organic fraction responsible for binding particles together. These practices favour pore continuity, improve aggregate and pore stability and, as a consequence, improve soil water retention and drainage.
- In many seasons and in much of the Australian wheat-sheep belt winter crops produce grain on stored soil water.
- Conditions which promote early vigour in crops promote canopy development and increase the proportion of stored soil water used for transpiration.
- Managing soils to promote the network of pores enhances the extent and depth of root system development and increases the soil's store of plant available water.
- Maximum yield for the majority of crops is achieved in the soil pH range of 5.5 – 6.5.
- Poor crop growth at pH extremes is due to one or more of the following:
  - hydrogen ion toxicity;
  - aluminium toxicity;
  - toxicity or reduced availability of nutrients;
  - detrimental effect of soil pH on organisms involved in the cycling of nutrients, particularly nitrogen.
- Management of pH extremes can be undertaken by:

- altering the pH with appropriate amendments;
- minimising practices which contribute to pH change;
- taking care with using fertilisers which alter soil pH.

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