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Forest Ecology and Management 220 (2005) 185–215
Forest Ecology and Management
www.elsevier.com/locate/foreco

The role of fire and nutrient loss in the genesis of the forest soils of Tasmania and southern New Zealand

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Abstract

The dominant soil patterns in forested or previously forested landscapes in southern New Zealand and Tasmania are described. Soil properties on adjacent sunny and shady aspects in hill country of the South Island of New Zealand are compared to soil properties under adjacent 'dry' and 'wet' eucalypt forest in Tasmania.

A soil contrast index or SCI is defined for comparing soil contrasts on parent materials of different absolute nutrient contents. Three soil groups are defined using the SCI. Group 1 soil pairs are stable New Zealand soils in which exchangeable Ca + Mg + K values are higher on drier sunny aspects than on moister shady aspects. Group 2 soil pairs are New Zealand soils in which soils on sunny aspects display evidence of topsoil erosion by wind; consequently some soil pairs on dry (sunny) aspects have lower levels of exchangeable Ca + Mg + K than soils on moister (shady) aspects. Group 3 soil pairs are Tasmanian. Soils on drier sites (under dry eucalypt forest) invariably have lower exchangeable Ca + Mg + K values than soils on moister sites (under wet eucalypt forest), which is the reverse of the pattern in SCI Group 1 soils in New Zealand.

Except on clay-rich parent materials, Tasmanian soils under dry forest generally have texture-contrast profiles and a mean C/N ratio in topsoils (A1 horizons) of 29. Soils under wet forest generally have uniform or gradational texture profiles and a mean topsoil C/N ratio of 15. The texture-contrast soils show strong clay eluviation with sand or sandy loam textures in upper horizons and clayey textures in lower horizons. However, in New Zealand texture-contrast soils are all but absent, and do not occur in the previously forested areas described in this paper. Topsoils (Ah horizons and soils sampled to 7.5 cm depth) in New Zealand areas sampled in this study have a mean C/N ratio of 15, regardless of whether they occur on sunny or shady aspects.

We propose that the frequency and spatial occurrence of fire are the dominant processes causing: (1) the marked difference in levels of nutrients and different topsoil C/N ratios in soils of Tasmania; (2) the development of texture-contrast soils under dry forests in Tasmania; and (3) the difference between soil patterns in New Zealand and Tasmania. Fire depletes nutrients in forests by causing losses to the atmosphere, losses by runoff, and losses by leaching. Nutrient loss by fire encourages fire-tolerant vegetation adapted to lower soil nutrient status, so frequent fire is a feedback mechanism that causes progressive soil nutrient depletion. By destroying organic matter and diminishing organic matter supply to the soil surface fire inhibits clay–organic matter linkages and soil faunal mixing and promotes clay eluviation. Fire frequency is likely to have increased markedly with the arrival of humans at ca. 34 000 years B.P. in Tasmania and ca. 800 years B.P. in New Zealand. We argue that texture-contrast soils

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doi:10.1016/j.foreco.2005.08.028

have not formed in New Zealand because of the short history of frequent fires in that country. A corollary of this conclusion is that texture-contrast soils in Tasmania are, at least in part, anthropogenic in origin.

Keywords: Burning; Fire; Texture-contrast soils; Australia; New Zealand; Landscape; Eucalypt forest; Clay eluviation; Aborigine

1. Introduction

The central and southern New Zealand South Island below the natural treeline, which occurs at approximately 1000–1200 m altitude, has been largely denuded of forest by fires of natural and human origin (Molloy et al., 1963; Molloy, 1964; McGlone et al., 1995; McGlone, 2001). The mean annual rainfall in this region, excluding the Central Otago semi-arid zone, is 500–1500 mm. The soils have remained generally stable and show an inherited soil pattern related to landscape position: the influence of topography (altitude and aspect), and by implication, microclimate, on soil properties in this area has been recognised since the early days of soil survey in New Zealand (Gibbs et al., 1945; Gibbs and Beggs, 1953; New Zealand Soil Bureau, 1968). These studies recognised trends of increasing leaching in soils with increasing altitude, and the different leaching status of soils on different aspects. The first detailed studies specifically relating soil chemical properties to aspect were conducted by Ives (1974), Ross (1971) and Cuff (1973) on soils of the previously forested Canterbury high country, and by Archer (1976) in the alpine and subalpine zones near Mt Cook. Later McIntosh et al. (1981, 1992, 2000) and Lynn et al. (2002) studied seasonally dry toposequences in more detail and confirmed the observations of overseas workers (e.g. Krauze et al., 1959; Cooper, 1960; Floate, 1965; Franzmeier et al., 1969; Losche et al., 1970), and of Ross (1971) and Cuff (1973) in New Zealand, that soils on sunny aspects tend to be less leached than soils on shady aspects at the same altitude, although local erosion can obscure this pattern.

Detailed studies detected the same leaching pattern in lowland hill country (McIntosh, 1986, 1992). While some of the New Zealand soils described show evidence of clay movement, either in the form of lamellae (e.g. McIntosh et al., 1992, Fig. 15), or as prominent clay coats in pores (McIntosh, 1986), clay percentage tends to be constant or to decrease with increasing soil depth in most soils, and texture-contrast soils like those occurring extensively in Australia (Isbell, 1996) are not present. The absence of texture-contrast soils in New Zealand is evident in the New Zealand Soil Classification (Hewitt, 1998), which contains no soil orders for texture-contrast soils,[1] whereas the Australian Soil Classification (Isbell, 1996) contains three.

The systematic study of forested and previously forested land in Tasmania commenced with the soil surveys of Laffan et al. (1995), Grant et al. (1995a) and Hill et al. (1995) in northern Tasmanian State forests. These survey results were summarised by Grant et al. (1995b) and indicated a strong association of forest types to soil classification (Laffan et al., 1998). Soils under ‘dry’ forest (eucalypt forests with an open heathy understorey) were commonly found to have texture-contrast (sandy over clayey textures) profiles as defined by Isbell (1996), whereas the soils under ‘wet’ forest (eucalypt forests with a dense broadleaved understorey, and rainforest) generally had uniform or gradational texture profiles (Isbell, 1996), The overall pattern was for soils under dry forest, which occurs predominantly in the drier climate zone and on drier sites such as sunny aspects in the wetter zone, to be more leached than soils under wet forest, which occurs predominantly in the wetter climate zone but also on shady aspects and in gullies in the drier zone.

Although there has been no detailed soil moisture monitoring in the above-described terrains, field observations, plant distribution (e.g. the widespread occurrence of drought-tolerant eucalypt species in dry forests), and common sense indicate that sunny slopes and dry forests have soils which are drier for longer than soils on shady slopes and under wet forests. It is evident therefore that there is contrasting soil

[1] Hewitt (1998) recognises clay illuviation features (e.g. clay skins, clay lamellae) at lower levels in the classification.

development in Tasmania and New Zealand. In Tasmania, under drier soil conditions, there is a strong tendency for texture-contrast soils to form, and for these drier soils to have fewer nutrients than soils which are more moist, but which have the same macroclimate. In contrast in New Zealand the drier soils show no tendency to form texture-contrast profiles, and unlike their counterparts in Tasmania, the drier soils tend to be less leached than moister soils nearby.

The above brief sketch of the soil patterns in southern New Zealand and Tasmania indicates that soil genesis in the two land masses at about the same latitude (41–46° south) appears to have proceeded along different pathways. The aim of the study described in this paper is to use the paired-site retrospective approach suggested by Raison et al. (1993) to investigate the differing forest soil distribution pattern between Tasmania and New Zealand in more detail, and to elucidate the processes causing the differing soil development pathways in the two land masses. The hypothesis being tested is that in terrain of the same rock type, given the same amount of time, detailed soil patterns (as demonstrated by described soil pairs) are determined by environmental influences such as microclimate, erosion and deposition, and that under similar environmental influences a similar soil pattern will result; conversely, that any dissimilarity of soil patterns between areas or land masses will mean that different environmental influences dominate.

2. Methods

This study investigates pedological and soil nutrient patterns in forested and previously forested soils of Tasmania and New Zealand. By pairing profiles on the same parent rocks and by expressing nutrient concentrations of profile pairs as ratios the effect of absolute differences of elemental composition between different parent rocks is minimised and comparisons between areas and countries is facilitated.

The detailed studies described are largely restricted to the medium-altitude high country of the South Island of New Zealand, the low altitude soils of the New Zealand Southland syncline, and the low and medium altitude forested land of Tasmania, as none of these areas, with the exception of one site pair noted in Table 1, had been fertilised at the time of soil sampling. Results are taken from various surveys originally conducted with different objectives, so there are some data gaps and differences in modes of presentation. Soil chemical analyses mostly followed the methods of Blakemore et al. (1987). Some analyses were by different methods, e.g. C analysis

Table 1
Profile chemistry and clay content of soil pair NZ1 from the tuffaceous greywacke terrain of the Southland syncline, New Zealand

Depth (cm)	Horizon	pH	Exchangeable cations				Total		Total P (mg/100 g)	P ret. (%)	Clay (%)
			Ca (cmol(+)/kg)	Mg (cmol(+)/kg)	K (cmol(+)/kg)	Ca + Mg + K (cmol(+)/kg)	C (%)	N (%)			
Kahiku soil (NZSC: Argillic Orthic Melanic soil) on sunny aspect											
0–24	Ah	5.8	15.4[a]	4.35	1.61	21.36	5.1	0.46	138[a]	38	38
24–35	AhBw	5.9	8.9	3.86	0.91	13.67	1.7	0.15	90	41	33
35–66	Bw1	5.8	8.7	3.55	0.36	12.61	1.0	0.12	91	40	40
66–92	Bw2	6.1	10.4	3.00	0.28	13.68	0.8	0.10	86	36	33
92–150	2Bt	6.3	13.8	4.19	0.30	18.29	0.5	0.07	66	32	33
Kaiwera soil (NZSC: Acidic Orthic Brown soil) on shady aspect											
0–18	Ah	5.6	9.9[a]	2.81	0.65	13.36	6.3	0.48	138[a]	60	39
18–30	AhBw	5.5	2.7	1.58	0.16	4.44	2.7	0.25	104	67	39
30–62	Bw1	5.3	0.9	0.95	0.08	1.93	1.7	0.15	91	71	40
62–90	Bw2	5.4	0.5	0.82	0.05	1.37	0.9	0.09	70	68	34
90–100	BC	5.4	0.5	0.17	0.08	1.47	0.2	0.02	52	58	22

Data from McIntosh (1988).
[a] Ah horizon total Ca and P values may have been increased by fertilisation.

for some profiles was by Walkley–Black digestion and for other profiles was by an ignition method. However, analyses for profile pairs were always by the same methods and therefore differences of analytical methodology are unlikely to affect interpretations or conclusions.

Soils were sampled from sites which showed no evidence of mass movement or soil mixing. As mentioned in Section 3, topsoils at some sites are likely to have been eroded by wind.

So that cation and organic matter differences between members of soil pairs (Tables 1–11 and 13) of widely differing absolute nutrient and carbon contents could be readily grouped and compared, for each site the log total C ratio (dry site/wet site) was plotted against log exchangeable Ca + Mg + K ratio (dry site/wet site). These plots allow soils to be grouped by their defining contrasting characteristics as a soil contrast index (SCI) (Fig. 3). Where full profile data were available, the SCI has been calculated for the same total profile depth for both profiles constituting a soil pair; where full profile data were not available, values for the A1 horizons (Tasmania) or Ah or 0–7.5 cm soils (New Zealand, depending on availability of data) were used. SCI values are expressed as sunny over shady (for the New Zealand soil pairs) or as dry forest over wet forest (for the Tasmanian soil pairs). Fig. 3 includes SCI values from the case study pairs described below as well as from additional sites not described in this paper.

3. Results—case studies

Twenty-four soil pairs are used in this study, 7 from Tasmania and 17 from New Zealand. To illustrate the predominant contrasts, the environmental and soil characteristics of selected soil pairs are described below. The soil classification for each soil of these pairs is given in Appendix 1.

3.1. New Zealand—the regional setting

The soil pairs described are from lowland Southland and the medium-altitude high country of inland South Canterbury (the Mackenzie hill country as defined by McIntosh and Hunter (1997)).

The Southland syncline in the far south of the South Island is formed from Triassic and Jurassic tuffaceous greywacke rocks (Wood, 1956; McKellar and Speden, 1978). Hard bands of sandstone and conglomerate rocks result in long ridges running northwest–southeast (Fig. 1). Altitudes vary from about 100 to 560 m a.s.l. and rainfall is in the range 1000–1400 mm. The soil parent material is slope colluvium which has been dated to 29 140 years B.P. at one location (McIntosh et al., 1990), but it is likely that slope instability continued to the close of the Last Glacial, i.e. to about 10 000 years B.P.

Until European pastoralists settled in Southland in the mid-1800s, most soils had podocarp-broadleaf forest cover, typically rimu (*Podocarpus cupressinum*), matai (*Prumnopitys taxifolia*), miro (*P. ferrugineus*), broadleaf (*Griselina littoralis*) and pepper tree (*Drimys colorata*). Such forest still covers higher altitude shady slopes. Broadleaf-podocarp forests are not adapted to fire and at the time of European settlement these forests may never have been burnt. Intense burns were used to clear land for grazing. Also present when the Europeans settled were red tussock (*Chionochloa rubra*) grasslands which resulted from Polynesians burning the previous forest cover after their arrival ca. 800 years B.P. (McGlone, 2001). These grasslands were probably occasionally re-burnt by the Polynesian population to maintain easy access and hunting grounds for the moa, possibly at intervals of a century or so.

Soils in the medium-altitude (400–1200 m a.s.l.) Mackenzie hill country (McIntosh and Hunter, 1997), are formed in stony colluvium, derived from underlying Triassic and Jurassic greywacke and subschist (Mutch, 1963), probably containing a minor but significant loess component, particularly in topsoils which tend to be less stony than subsoils and to have silt loam textures (McIntosh et al., 1992) typical of loess. The parent rocks are geosynclinal in origin and lack the tuffaceous component of the Southland syncline rocks. In the adjacent mid-Waitaki valley thermoluminescence and ^{14}C dating indicates that widespread erosion and deposition occurred in the period 70 000–10 000 years B.P. (Read et al., 1998). The parent materials in the Mackenzie hill country probably originated as screes in the Last Glacial (pre-10 000 years B.P.) when vegetation cover was patchy and dominated by *Chionochloa*, *Poa* and *Festuca* tussock grasses (McGlone et al., 1995).

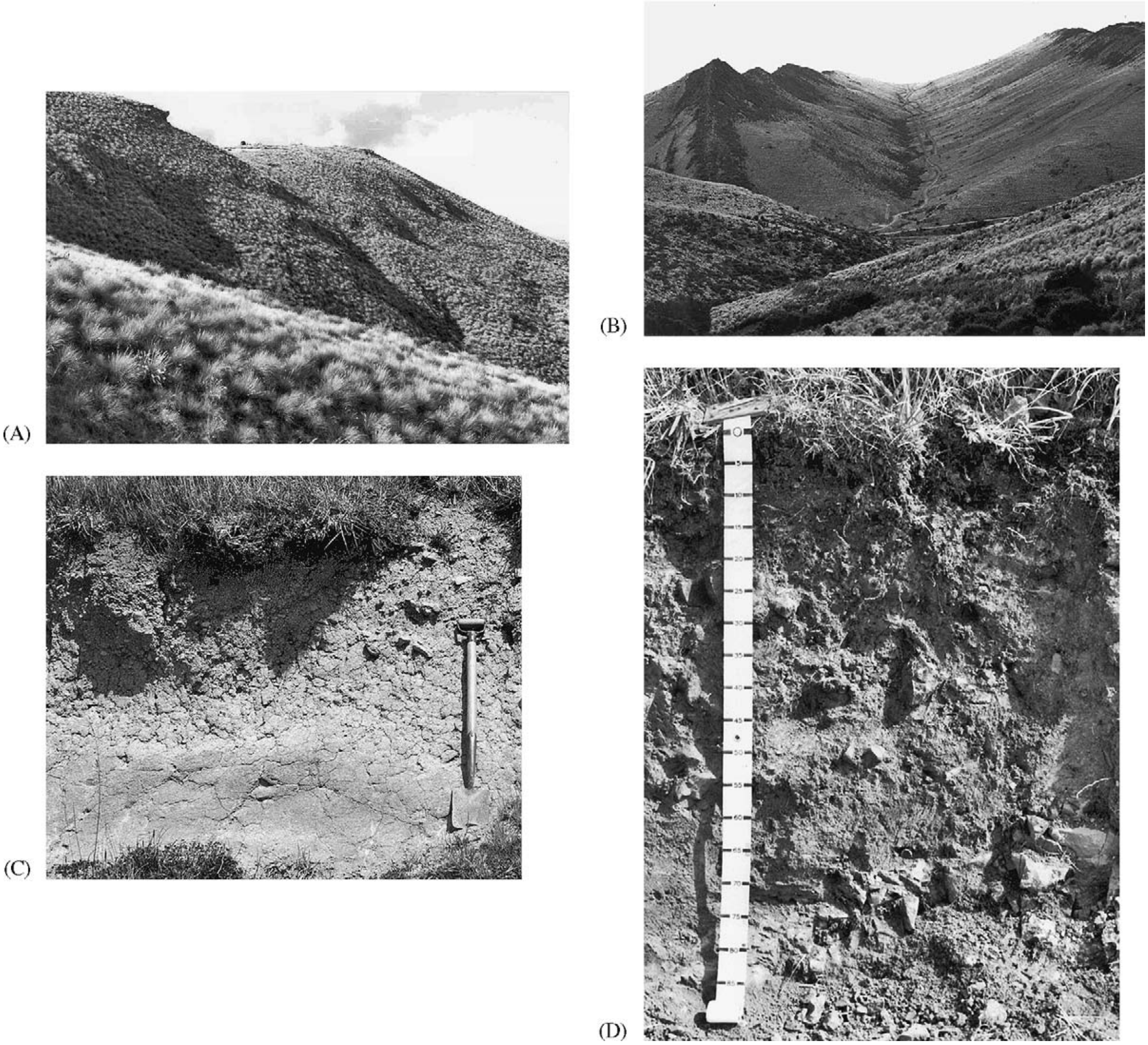


Fig. 1. (A) Low and mid-altitude hilly and steep land in New Zealand was forested until Polynesian-lit fires burnt most of the forest cover after about 800 years B.P. The forest vegetation was replaced by tussock grasslands (*Chionochloa* spp., *Poa* spp. and *Festuca* spp.). (B) The previously forested hills of the Southland syncline now have tussock grassland cover, where they have not been oversown and fertilised to establish improved pasture. (C) Kaihiku soils (NZSC: Argillic Orthic Melanic soils) with high base saturation occur on sunny slopes (right of B). (D) Kaiwera soils (NZSC: Acid Orthic Brown soils) are more leached of nutrients and occur on shady slopes (left of B). Neither Kaihiku nor Kaiwera soils show evidence of inter-horizon clay illuviation, although Kaihiku soils have prominent clay skins. See Appendix 1 for further soil classification details and Table 1 for soil analyses.

The evidence of forest remnants and subfossil remains (Molloy et al., 1963) indicates that in the Holocene (post-10 000 years B.P.) the soils were forested, with Hall's totara (*Podocarpus hallii*), celery pine (*Phyllocladus alpinus*), 'bog' pine (*Halocarpus bidwillii*), kowhai (*Sophora microphylla*) and scrub (e.g. *Discaria toumatou*, *Hymenanthera alpina*) on lower rainfall sites and beech forest (*Nothofagus menziesii*, *N. solandri*, and *N. solandri* var. *cliffordiodes*) on higher rainfall sites and at higher altitudes. The forest cover was modified in places by occasional natural fires and devastated over large areas by human-lit fires following the arrival of Polynesian settlers ca. 800 years B.P. (Molloy et al., 1963; McGlone et al., 1995; McGlone, 2001). Because fire-adapted tree species like eucalypts are absent from the New Zealand flora, tussock grasslands rather than regenerating trees replaced the forests. Consequently it can be assumed that most soils in the Mackenzie hill country have experienced few intense high-biomass fires but have experienced several low-intensity low-biomass grassland fires in the last 800 years. Since 1840 grassland fire frequency in tussock grasslands has increased because graziers burn the tussock cover regularly to stimulate the growth of succulent fresh tussock shoots.

Although local erosion has occurred as a result of fires and overgrazing by sheep and rabbits, initial theories of widespread soil instability following European settlement (Gibbs et al., 1945) have been shown to be unfounded (Whitehouse, 1984). The most prevalent form of erosion is probably wind erosion of topsoils on sunny slopes exposed to desiccating adiabatic northwesterly winds (McGowan et al., 1996; McGowan, 1997; Hewitt, unpublished information[2]) rather than widespread soil mixing. Paleosols in some areas indicate local periodic erosion (e.g. Molloy et al., 1963; McIntosh and Hunter, 1997, soil fact sheet 6), but the absence of buried soil layers or deep charcoal in most soil profiles and the predictability of soil patterns over large areas (McIntosh et al., 2000; Lynn et al., 2002) indicate the relative stability of the soils. We can assume that the soils retain much of their original character developed under several thousand years of forest vegetation and a much shorter time under grassland.

Broadly speaking, the Mackenzie hill country can be divided into three zones. Zone 1 is semi-arid (mean annual rainfall <500 mm); aspect has only a slight effect on soil leaching characteristics, because in most soils evaporation exceeds infiltration. Zone 2 is moister but seasonally dry and has a mean annual rainfall of 500–800 mm; aspect has an influence on soil profile development and nutrient status. Zone 3 is a moist high-altitude zone with mean annual rainfall >800 mm (extending into the subalpine and alpine regions above the former treeline), where detecting differences of leaching status between profiles on sunny and shady aspects is locally complicated by the effect of erosion (particularly above the former tree line), which causes soil loss and rejuvenation (McIntosh and Hunter, 1997, photo 2).

The soils described in this paper are taken from zone 2 and the lower-altitude (more stable) parts of zone 3, which are also the zones which supported podocarp or *Nothofagus* forests before natural or man-induced fires destroyed them. As mentioned above, because the forests were not adapted to regeneration after fire, the present soils may have experienced few fires (and possibly only one intense high biomass fire) during their Holocene development.

Within zone 2 and the lower part of zone 3 as defined above, the soil pattern has been established by McIntosh and Hunter (1997) and McIntosh et al. (1981, 1992, 2000). Soils at lower altitudes (400–750 m) within this zone, under a mean annual rainfall of about 500 mm (increasing with increasing altitude) generally have high base status (close to 100% base saturation) on both aspects, but are only slightly weathered (as deduced from phosphate retention values). Wind erosion of topsoils is likely on sunny slopes (see Footnote 2). At higher altitude (>750 m) mean annual rainfall probably exceeds 700 mm and here the greatest contrast between sunny and shady aspects occurs. Phosphate retention figures show that in this moister zone soils on shady slopes are more weathered than soils on sunny slopes. There is a tendency for soils on sunny slopes to have higher base saturation values than soils on shady slopes, indicating less leaching on sunny slopes, but the effect is not strongly expressed. At still higher altitudes (ca. 1000 m) where mean annual rainfall probably exceeds 800 mm, soil nutrient patterns are confused by active erosion, mainly on sunny slopes. Soils are less weathered than at lower altitudes (because of lower mean annual temperatures) and this difference is reflected in the lower P retention values of soils at higher altitudes (McIntosh et al., 2000).

A feature of all Mackenzie hill country soils is the lack of evidence of extensive clay eluviation. Some profiles have clay lamellae (for example, McIntosh et al., 1992, Fig. 15) but the process responsible for

[2] A.E. Hewitt, Landcare Research, Lincoln, New Zealand, in an unpublished report concerning erosion on hilly and steep land on Tara Hills High Country research station states: "Over the past 40 years, since the time of ^{137}Cs deposition, erosion has occurred in sunny aspect sites, exposed to north-west winds, on ridges and upper-convex back-slopes. Losses based on ^{137}Cs areal activity site means ranged from 20 to 32%, equivalent to losses in soil depth of 1.1–1.9 cm. Measurement of pedestals indicated that 47% of the erosion that formed the pedestals was prior to the 1950s".

these lamellae is intra-horizon rather than inter-horizon clay redistribution.

3.1.1. Soils of lowland Southland

3.1.1.1. Soil pair NZ1. The typical profiles for Kaihiku and Kaiwera soils (McIntosh, 1988, 1992), sampled on the opposing sides of the same ridge at 260 m altitude, demonstrate the marked effect of aspect on soil chemistry (Table 1). Total exchangeable cations (Ca + Mg + K) are present in much lower amounts in the shady aspect (Kaiwera) soil, and pH is also lower in this soil, indicating greater leaching than in the sunny aspect (Kaihiku) soil. In contrast, P retention and oxalate-extractable Al and Fe levels (not shown) are higher in the shady aspect soil, indicating more weathering and accumulation of poorly ordered allophanic clay minerals in this soil. The total C concentration is greater in the shady aspect soil, indicating a moister environment more favourable for organic matter accumulation. Total P values show no marked aspect effect.

Volume weight and stoniness measurements for all horizons of this soil pair (McIntosh, 1988) allow the leaching effect to be quantified for profiles to 100 cm depth: the more acid Kaiwera profile on the shady aspect has apparently ‘lost’ about Ca 18 t/ha, Mg 3 t/ha and K 2.2 t/ha relative to the Kaihiku profile on the sunny aspect, but total C, N and P contents of profiles show little or no difference. (Unknown replenishment from non-exchangeable sources of nutrients and other unquantified losses and gains means that true losses and gains cannot be calculated.)

McIntosh (1986, 1992) found that Kaihiku soils on sunny aspects had B horizon base saturation values in the range 55–81% and pH values in the range 5.6–6.8, in contrast to Kaiwera soils on shady aspects which had B horizon base saturation values in the range 7–46% and pH values of 4.8–5.7. There were marked pedological differences between the two soils: Kaihiku soils had prominent clay skins resulting from intra-horizon clay movement whereas Kaiwera soils lacked clay-illuviation features. In neither soil was a profile trend of clay percentage apparent (Table 1).

3.1.1.2. Soil pair NZ2. This higher-altitude (460 m) pair has Kaiwera soils on the sunny aspect and Venlaw soils on the shady aspect. The pair also demonstrates the greater acidity and greater leaching in the soil on the shady aspect in comparison with the soil on the sunny aspect (Table 2).

3.1.2. Soils of the moister Mackenzie hill country

3.1.2.1. Soil pair NZ3. In the moist western Mackenzie hill country (zone 3), soils mapped in the Tekoa set (New Zealand Soil Bureau, 1968) have developed under *Nothofagus* forest, under a mean annual rainfall of about 800–1000 mm. Many of these soils now have tussock grassland cover. Analyses (Table 3) demonstrate the lower pH and greater leaching of cations in the shady aspect soil, compared to the sunny aspect soil. The analysed soil pair does not follow the general pattern (McIntosh and Hunter, 1997) of having a higher Ah horizon C content in shady aspect soil.

3.1.2.2. Soil pair NZ4. A study on Longslip Station, predominantly in zone 3 as defined above, compared topsoil (0–7.5 cm) properties at 32 sites on either side of a ridge at altitudes of 730–1200 m (Table 4). The effect of aspect, and by implication microclimate, is most marked for cation values, which are lower in topsoils on shady aspects than on sunny aspects. P retention values are higher on shady aspects, which can be attributed to greater weathering on shady aspects than on sunny aspects. Total P, total C and total N values show few or no differences between aspects.

3.1.3. Soils of the drier Mackenzie hill country

3.1.3.1. Soil pair NZ5. An extensive topsoil survey of 72 sites on the Benmore Range (McIntosh et al., 2000) in the drier Mackenzie hill country (zone 2)

Table 2
Soil chemistry of soil pair NZ2 from the tuffaceous greywacke terrain of the Southland syncline, New Zealand

Soil	Classification (NZSC)	Aspect	pH	Exchangeable cations			Total	
				Ca (cmol(+)/kg)	Mg (cmol(+)/kg)	K (cmol(+)/kg)	C (%)	N (%)
Kaiwera	Acidic Orthic Brown soil	Sunny	5.1–5.5	4.57	2.72	1.39	3.3	N.D.
Venlaw	Acidic Allophanic Brown soil	Shady	4.7–5.2	0.52	0.63	0.33	7.8	0.34

Values shown are mean values for horizons to 64 cm depth, weighted for horizon thickness. Data from McIntosh (1988). N.D.: not determined.

Table 3
Profile chemistry of soil pair NZ3 from the greywacke terrain of the Mackenzie hill country, New Zealand

Depth (cm)	Horizon	pH	Exchangeable cations				Total		P ret. (%)	Clay (%)
			Ca (cmol(+)/kg)	Mg (cmol(+)/kg)	K (cmol(+)/kg)	Ca + Mg + K (cmol(+)/kg)	C (%)	N (%)		
Tekoa "sunny" (NZSC: Allophanic Brown soil)										
0–13	Ah	5.4	3.57	0.89	0.86	5.35	4.7	0.24	36	N.D.
13–28	AhBw	5.2	0.92	0.26	0.43	1.61	2.3	0.12	41	N.D.
28–40	Bw	5.2	0.35	0.11	0.33	0.79	2.0	0.10	60	N.D.
40–57	2BC	5.3	0.25	0.03	0.09	0.37	2.1	0.09	87	N.D.
Tekoa "shady" (NZSC: Allophanic Brown soil)										
0–9	Ah	4.7	1.66	0.66	0.49	2.81	4.0	0.24	47	N.D.
9–25	AhBw	4.8	0.83	0.22	0.14	1.19	2.1	0.12	59	N.D.
25–55	Bw	5.0	0.23	0.04	0.06	0.33	1.9	0.08	64	N.D.
55–85	BC	5.1	0.04	0.01	0.05	0.43	1.1	0.05	47	N.D.

Data from McIntosh and Hunter (1997). N.D.: not determined.

showed that Ah horizon C and N values (Table 5) are strongly affected by aspect (on an area basis the mean aspect difference is C 37 t/ha and N 2.3 t/ha, when Ah horizon thickness and volume weight are taken into account) but cation values are only slightly affected (K and Mg) or unaffected (Ca). Combining the evidence of the Ah horizon data and profile data (not tabulated), it appears that there are two processes acting in the mid-altitude Mackenzie hill country: the predominant subsoil process is a tendency for leaching and loss of cations to be greater on shady aspects, as to be expected from the generally wetter state of these soils resulting from lower evapotranspiration losses. Acting to obscure this tendency for profiles on shady aspects to be more leached is the susceptibility of Ah horizons on sunny aspects to become eroded by wind, which depletes available nutrients in Ah horizons on sunny aspects.

3.1.3.2. Soil pair NZ6. In the southern Benmore Range, McIntosh, Hewitt and Burgham (Landcare Research, unpublished data) described and analysed profiles at 895 m on a sunny aspect and at 810 m on a shady aspect, in steep country with slopes of 35° (Table 6). These profiles show typical soil patterns: total C and N are lower on sunny aspects than on shady aspects, but weathering, as indicated by P retention values, is greater in upper horizons on shady aspects, which also have higher oxalate-extractable Al and Fe values (not shown). Values of total exchangeable cations are similar on both sunny and shady aspects. The lack of an aspect-related difference of exchange-

Table 4
Topsoil (0–7.5 cm) chemistry of soil pair NZ4, consisting of multiple sites on either side on a greywacke and subschist ridge on Longslip Station, Mackenzie hill country, New Zealand

	Sunny aspect ($n = 16$)	Shady aspect ($n = 16$)	S.E.D.[a]
Total C (%)	3.1	3.3	0.35
Total N (%)	0.26	0.26	0.026
P ret. (%)	26	42	2.4
Total P (g/t)	839	816	51
Exch. Ca (cmol(+)/kg)	4.8	2.5	0.49
Exch. Mg (cmol(+)/kg)	0.99	0.57	0.10
Exch. K (cmol(+)/kg)	1.02	0.77	0.08

Data from McIntosh et al. (1981).
[a] Standard error of the difference between means.

Table 5
Soil chemistry of soil pair NZ5, consisting of multiple sites on the greywacke terrain of the Benmore Range, Mackenzie hill country, New Zealand

	Sunny aspect ($n = 36$)	Shady aspect ($n = 36$)	LSD ($P < 0.05$)
Total C (%)	1.94	3.53	0.35
Total N (%)	0.14	0.24	0.02
P ret. (%)	20	34	3
Exch. Ca (cmol(+)/kg)	4.08	3.92	0.60
Exch. Mg (cmol(+)/kg)	0.82	0.70	0.12
Exch. K (cmol(+)/kg)	0.55	0.44	0.09

Data from McIntosh et al. (2000).

Table 6
Profile chemistry and clay content of soil pair NZ6 from the greywacke terrain of the Benmore Range, Mackenzie hill country, New Zealand

Depth (cm)	Horizon	pH	Exchangeable cations				Total (%)		P ret. (%)	Clay (%)
			Ca (cmol(+)/kg)	Mg (cmol(+)/kg)	K (cmol(+)/kg)	Ca + Mg + K (cmol(+)/kg)	C (%)	N (%)		
Typic Orthic Recent soil on sunny aspect										
0–15	Ah	5.6	4.6	1.11	0.49	6.20	1.45	0.13	15	15
15–25	AhBw	6.0	3.7	0.96	0.29	4.95	0.79	0.08	17	13
25–41	Bw1	6.1	3.3	0.98	0.25	4.53	0.76	0.07	21	11
41–52	Bw2	6.1	3.2	0.98	0.19	4.37	0.36	0.04	14	10
52–70	BC	6.3	2.8	0.83	0.10	3.73	0.16	0.02	11	6
Typic Immature Pallic soil on shady aspect										
0–15	Ah1	5.0	6.3	1.68	0.59	8.57	3.70	0.27	38	16
15–31	Ah2	5.4	2.4	0.59	0.14	3.13	1.70	0.12	43	10
31–53	Bw	5.8	1.7	0.38	0.09	2.17	0.61	0.06	27	4
53–70	BC1	5.9	3.0	0.75	0.11	3.86	0.27	0.04	18	7
70–92	BC2	6.2	6.1	1.27	0.18	7.55	0.19	0.03	14	9

Data from McIntosh, Hewitt and Burgham (Landcare Research, unpublished).

able cations in these dry soils is attributed to the fact that they are only slightly leached as they border on the semi-arid climate zone.

3.2. Tasmania—the regional setting

Most soils in eastern Tasmania are formed in one of four parent rocks: Silurian-Devonian greywacke (Mathinna beds), Devonian granite/adamellite, Devonian granodiorite, or Jurassic dolerite. Quartz is a major component of the first three rocks, but in dolerite quartz is either a very minor component, or absent. As in New Zealand, periglacial conditions during the Last (Rowallan) Glacial probably rejuvenated soils, either by eroding pre-existing soils to weathered or unweathered parent rock, or by causing accumulation of fresh slope deposits. Sigleo and Colhoun (1982), referring to Last Glacial lowland deposits in southern Tasmania, remarked that ‘‘the dolerite . . . and the sandstone slope deposits, including aeolian lenses . . . suggest that the cool/cold, dry conditions of sandsheet accumulation were preceded by cool/cold, moist conditions when heavy rain, snowmelt and strong frosts facilitated sheetwash and debris movement on slopes’’. Aeolian deposits in the Huon valley dated to about 28 000 years B.P. (McIntosh et al., 2004) attest to the dry windy conditions and probable extensive bare ground and non-forest vegetation of herbs and shrubs (Colhoun et al., 1994) during this time. Local deep red weathering in some soil parent materials (Grant et al., 1995b) indicates that they were probably weathered in an earlier interglacial than the present Holocene epoch; these older parent material have not been totally stripped from the landscape. Pollen studies (Markgraf et al., 1986; Colhoun et al., 1994) indicate that the transition from a grassy shrubland (glacial) landscape to a forested (interglacial) more stable landscape occurred about 12 000 years B.P.

A feature of Tasmanian soils is the marked contrast between soils under ‘dry’ forest and soils under ‘wet’ forest (Laffan et al., 1998). Dry sclerophyll forest is open eucalypt forest composed of drought-tolerant species adapted to understorey fire (Duncan, 1996). Typically the species most represented on lowlands are *Eucalyptus amygdalina* and *E. obliqua* (*E. sieberi* in northeast Tasmania) trees, an understorey of prickly shrubs (Epacridaceae) and bracken (*Pteridium esculentum*) (Fig. 2A), and commonly a high proportion of bare ground. In contrast wet sclerophyll forest is eucalypt forest composed of species that are ill-adapted to understorey fire (Duncan, 1996). Typically the species most represented is *E. regnans* (with *E. obliqua*) with a dense understorey of broadleaved shrubs such as dogwood (*Pomaderris apetala*), musk (*Olearia argophylla*) and stinkwood (*Zieria arborescens*) (Fig. 2D).

Fires are more frequent, but generally less intense (of shorter duration and sometimes cooler) in dry forests than in wet forests (Duncan and Brown, 1995).

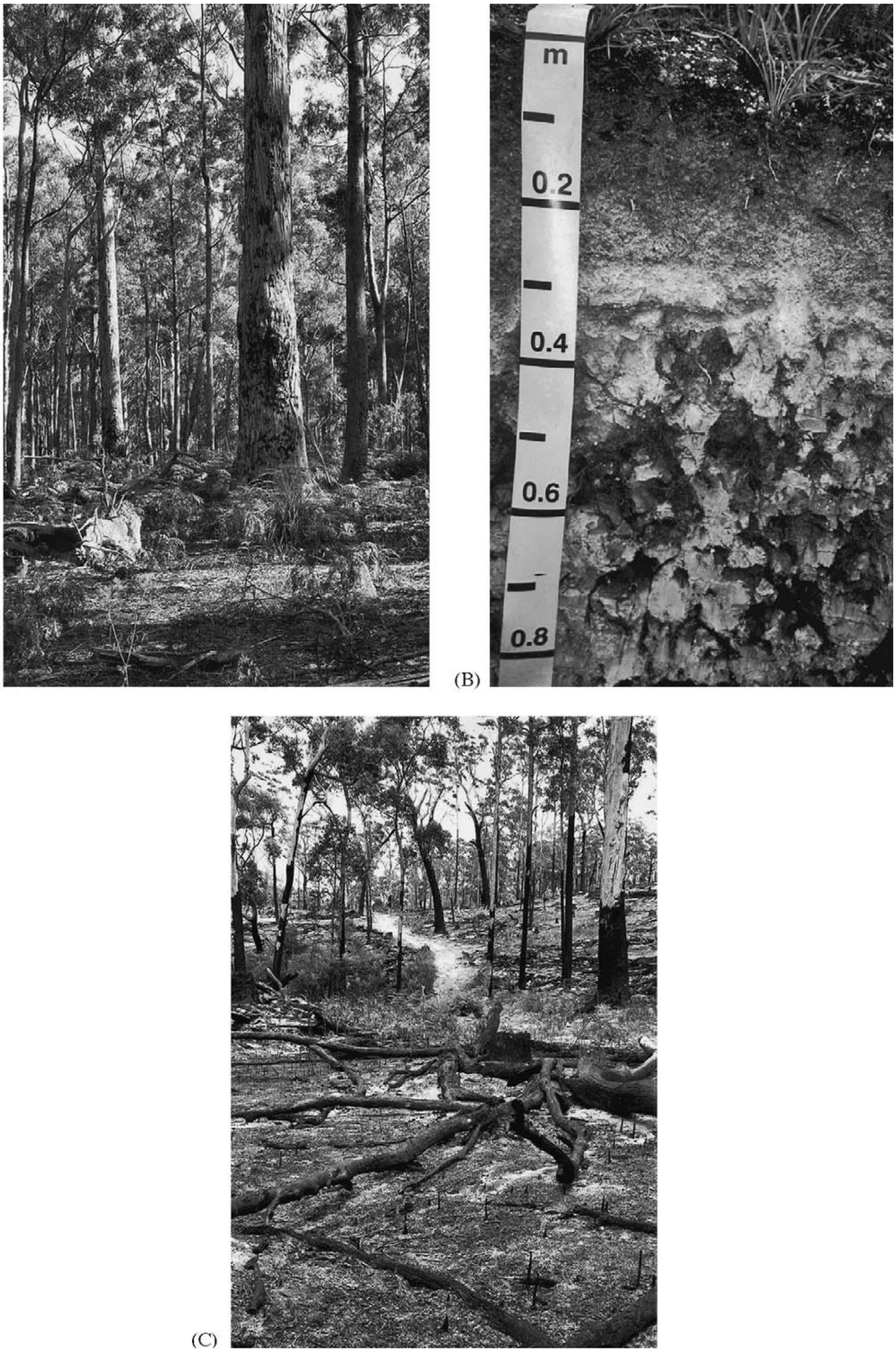
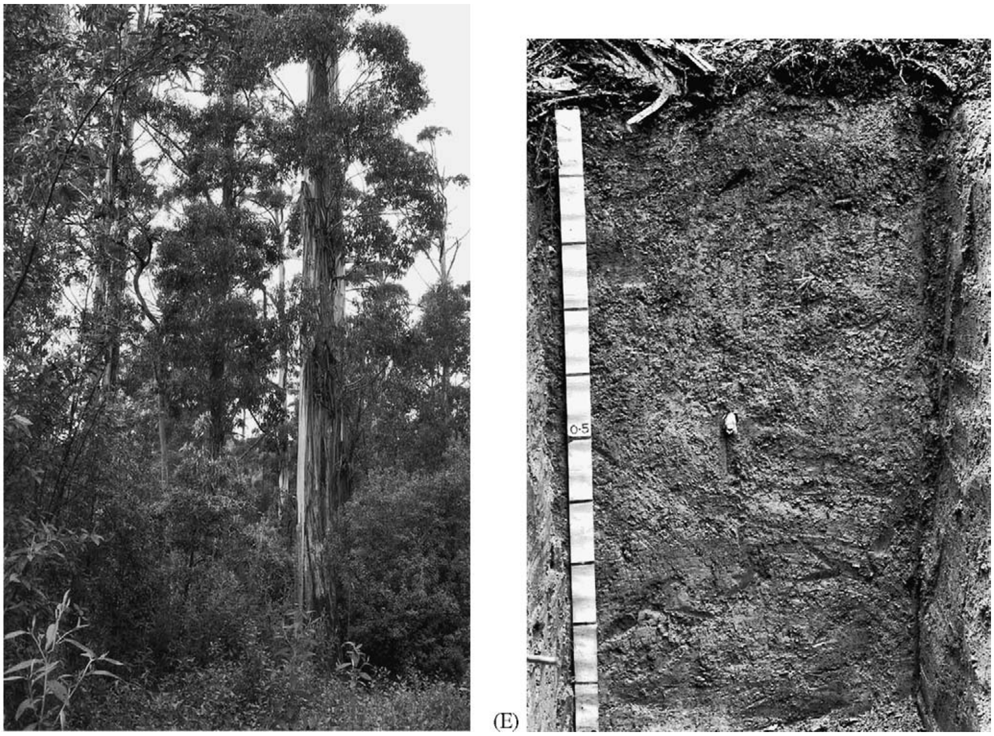


Fig. 2. (A) Typical Tasmanian dry forest (*E. obliqua* and *E. amygdalina* with bracken (*P. esculentum*)), on the texture-contrast Mckay soil (ASC: Kurosol) developed in granodiorite in northeast Tasmania. (B) Similar dry forest grows on the texture-contrast Jensen soil (ASC: Kurosol)



(D) (E)

Fig. 2. (*Continued*).

The understorey and litter in dry forests is commonly burnt, but in these understorey fires many canopy trees survive (Fig. 2). However, in occasional intense fires in dry forests all vegetation will be burnt. The distribution of the two forest types is linked to regional climate—the dry forest types occur in the drier areas (mean annual rainfall <1000 mm), and the wet forest types occur in areas with higher mean annual rainfall. However, wet forest frequently occurs on shady aspect hill slopes and in gullies where mean annual rainfall is much less than 1000 mm, and dry forest can occur in high rainfall areas. The transition between the two forest types is generally sharp.

Laffan et al. (1998) suggested that frequent fires might explain the association of forest types with soils in northeast Tasmania and described the soil-vegetation pattern, which can be generalised as follows: where contrasting forest types occur together on soils containing significant quartz (i.e. all soils except those formed in dolerite), dry forest types are generally associated with texture-contrast soils (Kurosols or Chromosols, which approximate to Haplustults in USDA Soil Taxonomy (USDA, 1998)). In contrast wet forest types are normally associated with uniform texture or gradational texture soils (Dermosols, which approximate to Hapludults and Dystrudepts in USDA Soil Taxonomy (USDA, 1998)). On dolerite, soils under wet forest types have uniform or gradational texture profiles (Eutrophic Ferrosols or Eutrudepts) whereas under dry forest the soils include both texture-contrast profiles (Chromosols) and uniform or gradational profiles (Mesotrophic Ferrosols or Dystrudepts).

developed in granite; note the sharp change in texture from sandy horizon (ca. 4% clay) above 33 cm depth to clay-rich horizons (>50% clay) below this depth. (C) Natural fires are frequent in this type of forest: the understorey is totally burnt (except around streams), heavier debris such as fallen trees is partly burnt; most large trees survive fires when the understorey fuel load is low. (D) In contrast wet forest on the gradational Blumont soil (ASC: Dermosol; scale in 10 cm intervals) (E) is dominated by *E. regnans* and has a dense broadleaf understorey. Such wet forest sites are very seldom burnt. See Appendix 1 for further soil classification details and Tables 8 and 9 for soil analyses.

All the examples of soil pairs below are taken from eastern Tasmania, at altitudes ranging between 100 and 400 m a.s.l. and under a mean annual rainfall of 700–1200 mm. With the exception of the Blumont soil (soil pair 5.1, Fig. 2 and Appendix 1) all Tasmanian soils described had maximum exchangeable sodium percentages in subsoil horizons in the range 1.5–5.3% (mean value 2.5%). Sodicity is therefore unlikely to be a factor in the development of the soils described. The Blumont soil (an Acidic Mesotrophic Brown Dermosol) had an exchangeable Na value of 1.37 cmol(+)/kg in the B3t horizon and an exchangeable sodium percentage in this horizon of 14.6%. The significance of this value is discussed further under the heading 'Soil pair Tas3' below.

3.2.1. Soils in sandstone (Mathinna beds)

Various soils formed in Silurian-Devonian sandstone (Mathinna beds) have been mapped in northern Tasmania (Laffan et al., 1995). The Retreat soils under dry forest and Maweena soils under wet forest have been chosen for comparison as they occur under similar macroclimate.

3.2.1.1. Soil pair Tas1. Apart from the very marked pedological difference between the texture-contrast Retreat soil under dry forest and the gradational Maweena soil under wet forest, there are marked contrasts of chemical properties between these two soils (Table 7), viz.: (1) lower total C values under dry forest compared to wet forest; (2) lower total Ca + Mg + K values under dry forest; (3) lower total P values in soils under dry forest. The two soils have contrasting pH trends. The texture-contrast soil under dry forest has lowest pH in the surface horizons, whereas the gradational soil under wet forest has the reverse trend.

On an area basis to 80 cm depth the dry forest soil contains 19 t/ha less total C than the wet forest soils (81 t/ha versus 100 t/ha), 3 t/ha less total N (5.0 t/ha versus 8.0 t/ha) and 91 t/ha less total P (62 t/ha versus 153 t/ha). The incomplete cation analyses do not allow accurate calculation of cation differences, but the subsoil analyses available indicate that subsoils under dry forest contain about half the cation levels of subsoils under wet forest.

3.2.2. Soils in granite and granodiorite

On granite and adamellite the soils have been mapped (Grant et al., 1995a) as Jensen soils under dry forest and Stronach soils under wet forest, and on granodiorite as Mckay soils under dry forest and Blumont soils under wet forest.

3.2.2.1. Soil pair Tas2. In granite terrain the primary contrast is between Jensen soils under dry forest and Stronach soils under wet forest (Table 8). The Jensen soil has extreme texture-contrast characteristics. The

Table 7
Profile chemistry and clay content of soil pair Tas1 on sandstone (Mathinna beds) in northeast Tasmania

Depth (cm)	Horizon	pH	Exchangeable cations				Total		Total P (mg/kg)	P ret. (%)	Clay (%)
			Ca (cmol(+)/kg)	Mg (cmol(+)/kg)	K (cmol(+)/kg)	Ca + Mg + K (cmol(+)/kg)	C (%)	N (%)			
Retreat soil (ASC: Brown Kurosol) under dry forest											
0–13	A1	4.7	2.00	1.30	0.33	3.63	1.1	0.12	55	7	8
13–24	A21	4.8	N.D.	N.D.	N.D.	–	1.5	0.06	32	3	4
24–35	A22e	5.1	N.D.	N.D.	N.D.	–	0.6	0.03	18	12	4
35–41	B1t	4.9	N.D.	N.D.	N.D.	–	0.6	0.03	42	27	24
41–83	B2t	5.3	0.49	0.10	0.10	0.69	0.5	0.03	89	55	60
83–95	BC	5.4	N.D.	N.D.	N.D.	–	0.4	N.D.	95	43	36
Maweena soil (ASC: Brown Dermosol) under wet forest											
0–14	A1	5.2	2.70	1.60	0.39	4.69	2.4	0.19	179	12	18
14–26	AB	4.9	N.D.	N.D.	N.D.	–	0.9	0.06	104	15	21
26–36	B21t	4.9	N.D.	N.D.	N.D.	–	0.5	0.05	91	21	24
36–58	B22t	4.8	0.56	0.88	0.27	1.71	0.5	0.04	142	24	20
58–80	B23t	4.8	N.D.	N.D.	N.D.	–	0.5	0.05	202	31	24

Data from Laffan et al. (1995). N.D.: not determined.

Table 8
Profile chemistry and clay content of soil pair Tas2 on granite in northeast Tasmania

Depth (cm)	Horizon	pH	Exchangeable cations				Total		Total P (mg/kg)	P ret. (%)	Clay (%)
			Ca (cmol(+)/kg)	Mg (cmol(+)/kg)	K (cmol(+)/kg)	Ca + Mg + K (cmol(+)/kg)	C (%)	N (%)			
Jensen soil under dry forest (ASC: Yellow Kurosol)											
0–9	A1	4.9	1.5	0.84	0.13	2.47	2.92	0.079	43	0	N.D.
9–20	A21	4.8	N.D.	N.D.	N.D.	–	1.01	0.036	28	0	4
20–32	A22	4.7	N.D.	N.D.	N.D.	–	0.23	0.007	20	0	4
32–52	B21t	5.4	1.4	1.4	0.18	2.98	0.79	0.032	44	15	52
52–69	B22t	5.5	N.D.	N.D.	N.D.	–	1.01	0.043	81	57	76
69–110	B23t	5.1	N.D.	N.D.	N.D.	–	0.9	0.029	84	42	82
Stronach soil under wet forest (ASC: Brown Dermosol)											
0–12	A11	5.9	12.0	2.3	0.75	15.05	5.27	.0.313	348	21	26
12–37	A12	5.6	N.D.	N.D.	N.D.	–	3.04	0.209	246	23	22
37–50	B1	5.3	N.D.	N.D.	N.D.	–	2.43	0.133	201	33	34
50–95	B2t	4.8	0.81	0.59	0.24	1.64	1.71	0.123	195	45	40
95–120	BC	4.6	N.D.	N.D.	N.D.	–	1.13	0.063	131	45	52

Data from Grant et al. (1995a). N.D.: not determined.

A21 and A22 horizons contain 92–96% sand and 4% clay and consequently their P retention values are zero. In contrast the B22t horizon contains 19% sand and 76% clay and has a P retention value of 57%. The profile has lower C, N and P levels than the Stronach profile. The texture-contrast Jensen soil has low pH and low total cation values at or near the soil surface, whereas the gradational Stronach soil has high total exchangeable cation and P levels in the A1 horizon and pH increases towards the soil surface, as would be expected from a long history of nutrient cycling and accumulation under the wet forest.

3.2.2.2. Soil pair Tas3. The Mckay-Blumont soil pair (Table 9) formed in granodiorite parent rocks shows very similar characteristics to the Jensen–Stronach pair, viz.: a texture-contrast profile under dry forest but a gradational profile under wet forest. P retention values are very low in the A1 and A2 horizons of the Mckay soil, and these horizons have loamy sand and

Table 9
Profile chemistry of soil pair Tas3 on granodiorite in northeast Tasmania

Depth (cm)	Horizon	pH	Exchangeable				Total		Total P (mg/kg)	P ret. (%)	Clay (%)
			Ca (cmol(+)/kg)	Mg (cmol(+)/kg)	K (cmol(+)/kg)	Ca + Mg + K (cmol(+)/kg)	C (%)	N (%)			
Mckay soil under dry forest (ASC: Yellow Kurosol)											
0–12	A1	4.4	1.25	0.52	0.11	1.88	3.20	0.08	34	1	N.D.
12–30	A2e	4.8	0.46	0.23	0.04	0.73	0.62	0.04	21	2	N.D.
30–46	B2	5.0	0.59	0.56	0.11	1.26	0.31	0.02	17	35	N.D.
46–92	B2t1	4.9	0.68	1.32	0.29	2.29	0.48	0.03	23	26	N.D.
92–110	B2t2	5.0	0.44	1.28	0.30	2.02	0.22	0.02	26	23	N.D.
Blumont soil under wet forest (ASC: Brown Dermosol)											
0–16	A1	5.4	3.97	1.53	0.59	6.09	2.74	0.21	156	25	N.D.
16–43	B1	5.2	1.96	1.09	0.28	3.33	1.30	0.10	95	26	N.D.
43–77	B2	5.2	1.51	0.91	0.27	2.69	1.01	0.07	70	29	N.D.
77–100	B3t	4.9	0.81	0.73	0.32	1.86	0.48	0.03	39	29	N.D.

Data from Grant et al. (1995a). N.D.: not determined.

sandy loam texture, respectively. The Mckay soil also has lower overall C, N, P and exchangeable cation levels relative to the Blumont soil. The Mckay soil has a trend of decreasing pH towards the soil surface, whereas the opposite trend occurs in the Blumont soil. Exchangeable cation and total P values also increase towards the soil surface in the Blumont soil. The high exchangeable sodium percentage in the B3t horizon of Blumont soil is unusual as other Tasmanian soils described have a mean maximum exchangeable sodium percentage of 2.5%. The value is not associated with highly dispersible soil material, as the percentage of water-stable aggregates (Laffan et al., 1996) in this horizon is 52% and soil structure is moderately developed. It is considered that the exceptional exchangeable Na values are likely to be caused by weathering of plagioclase-rich granodiorite in this soil.

3.2.3. Soils in dolerite

The forest pattern on Jurassic dolerite is similar to that on greywacke, granite and granodiorite parent rocks. Dry forest (e.g. *Eucalyptus obliqua* and *E. amygdalina* with an understorey of heath and bracken) occurs more frequently on sunny aspects, and wet forest (e.g. *E. regnans*, *E. globulus* with an understorey of *Pomaderris apetala* and *Bedfordia salicina*) occurs mostly under wetter climates and on shady aspects.

3.2.3.1. Soil pair Tas4. The two soils selected for comparison both have gradational profiles with red subsoils, and occur close to each other in northeast Tasmania (Table 10). Holloway soils occur under dry forest and Excalibur soils under wet forest (Laffan et al., 1995). Holloway soils have an A2 horizon that is either only weakly developed or absent; clay percentage increases gradually from 17% in the A1 horizon to 47% in the lower B2 horizon. Excalibur soils have a similar texture trend: clay percentage increases from 20% in the A1 horizon to 48% in the lower B2 horizon. However, Holloway soils have lower exchangeable cation levels, lower total C and N levels in the A1 horizon, and lower total P values throughout profiles, than Excalibur soils (Table 10), which results in Holloway soils being classified as Mesotrophic Ferrosols (Dystrudepts) whereas Excalibur soils are classified as Eutrophic Ferrosols (Eutrudepts) (Appendix 1).

3.2.3.2. Soil pair Tas5. In southeast Tasmania similar soils to Holloway and Excalibur soils have been described and named as Bream and Wielangta soils (McIntosh et al., 2001a,b). The contrasts between the properties of this soil pair (Table 11) are greater than for soil pair Tas4. In particular, total P concentrations in the Bream soil under dry forest are about one-third those under wet forest, and the A1 horizon of the Bream soil is very thin (3 cm).

Another contrasting soil pair on dolerite consists of the widespread texture-contrast Eastfield soil under dry forest and the gradational Murdunna soil under wet forest (Grant et al., 1995b). This soil pair has not been included in this study because Murdunna soils have not been fully analysed and there is some doubt as to

Table 10
Profile chemistry and clay content of soil Tas4 on dolerite in northeast Tasmania

Depth (cm)	Horizon	pH	Exchangeable cations				Total		Total P (mg/kg)	P ret. (%)	Clay (%)
			Ca (cmol(+)/kg)	Mg (cmol(+)/kg)	K (cmol(+)/kg)	Ca + Mg + K (cmol(+)/kg)	C (%)	N (%)			
Holloway soil (ASC: Mesotrophic Ferrosol) under dry forest											
0–19	A11	N.D.	0.99	0.55	0.24	1.78	1.5	0.06	76	22	17
19–39	A12	N.D.	N.D.	N.D.	N.D.	–	0.8	0.03	74	20	26
39–66	B21t	N.D.	N.D.	N.D.	N.D.	–	0.1	0.02	81	29	34
66–124	B22t	N.D.	1.70	2.50	0.18	4.38	0.1	0.01	87	36	47
Excalibur soil (ASC: Eutrophic Ferrosol) under wet forest											
0–15	A1	5.9	7.20	2.70	0.80	10.7	4.1	0.21	195	64	20
13–47	B1	5.9	N.D.	N.D.	N.D.	–	1.1	0.07	145	60	28
47–83	B21t	6.1	3.00	3.20	0.53	6.73	0.4	0.04	158	52	36
83–120	B22t	6.1	N.D.	N.D.	N.D.	–	0.4	0.03	128	58	48

Data from Laffan et al. (1995). N.D.: not determined.

Table 11
Profile chemistry of soil pair Tas5 on dolerite in southeast Tasmania

Depth (cm)	Horizon	pH	Exchangeable cations				Total		Total P (mg/kg)	P ret. (%)	Clay (%)
			Ca (cmol(+)/kg)	Mg (cmol(+)/kg)	K (cmol(+)/kg)	Ca + Mg + K (cmol(+)/kg)	C (%)	N (%)			
Bream soil (ASC: Mesotrophic Ferrosol) under dry forest											
0–3	A1	6.0	11.85	3.09	0.58	15.52	6.33	0.21	154	51	N.D.
3–35	B1	5.3	1.21	1.63	0.22	3.06	1.61	0.06	79	53	N.D.
35–60	B2	5.3	1.55	3.66	0.47	5.68	0.91	0.05	71	59	N.D.
60–90	B3	5.3	1.99	7.88	0.40	10.27	0.65	0.04	73	66	N.D.
Wielangta soil (ASC: Eutrophic Ferrosol) under wet forest											
0–13	A1	5.6	8.16	3.67	1.46	13.29	5.71	0.32	444	50	N.D.
13–28	AB	5.5	4.13	3.70	1.37	9.20	2.20	0.15	358	57	N.D.
28–57	B21	5.3	4.12	4.33	1.09	9.54	1.60	0.10	216	54	N.D.
57–110	B22	5.5	5.12	5.65	1.25	12.02	0.59	0.04	215	51	N.D.

Data from McIntosh et al. (2001a,b). N.D.: not determined.

whether the sandy upper horizons of Eastfield soils are wholly dolerite-derived: aeolian sands may contribute to the parent material at some locations (R. Osak, University of Tasmania, personal communication).

3.3. Summary of soil pair differences

As cation and organic matter differences have been shown to be key differentiae between members of soil pairs (Tables 1–11), the ratios of exchangeable Ca + Mg + K and total C for soil pairs have been calculated for each soil pair and plotted against each other (Fig. 4) as soil contrast indices (SCIs), as described in Section 2. The SCI values fall into three distinct groups (Fig. 3): Group 1, consisting of stable New Zealand soils, from the lowlands and wetter Mackenzie hill country; Group 2, consisting of soils from the drier Mackenzie hill country showing various degrees of erosion on sunny slopes; and Group 3, consisting of Tasmanian soils.

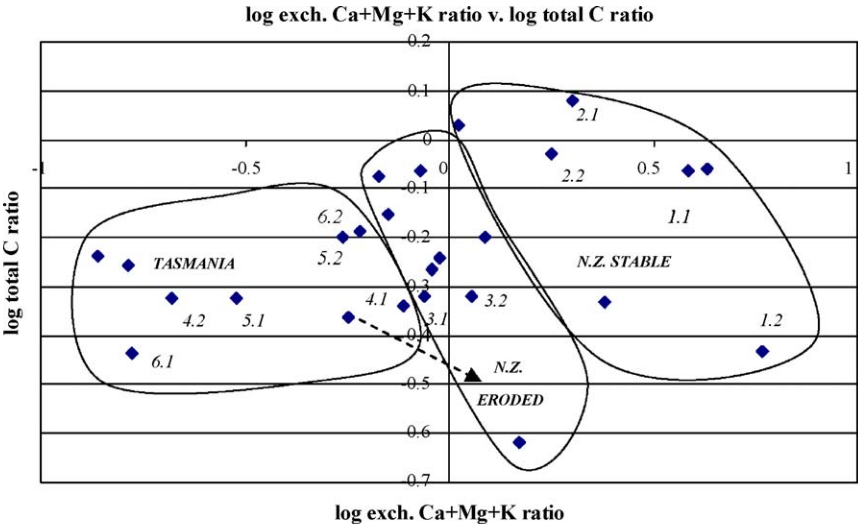


Fig. 3. Plot of soil contrast index or SCI (log total C ratio vs. log exch. Ca + Mg + K ratio) for soil pairs described in the text (case study numbers in italics) and other undescribed soil pairs. The soils are grouped into stable New Zealand soils (Group 1), New Zealand soils in which the sunny aspect soil of each soil pair shows signs of topsoil erosion (Group 2) and Tasmanian soils (Group 3). The dotted arrow indicates an outlying Group 2 soil pair which plots among Group 3 soils.

Table 12
C/N ratios of A horizons or topsoils (0–7.5 cm) of soils making up the pairs used in Fig. 3

Source	C/N ratio
New Zealand sunny aspect ($n = 16$)	14.9 (3.1)
New Zealand shady aspect ($n = 16$)	14.9 (3.8)
Tasmanian dry forest ($n = 7$)	28.5 (10.3)
Tasmanian wet forest ($n = 7$)	14.9 (3.5)

Sample standard deviations are in parenthesis.

A negative log ratio for C indicates that the sunny or dry forest profile contains less C than its shady or wet profile pair. A negative log ratio for exchangeable Ca + Mg + K indicates that the sunny or dry forest profile contains fewer total cations than its paired shady or wet forest profile pair.

In addition, the C/N ratios of A1 horizons (Tasmania) or Ah horizons and 0–7.5 cm soils (New Zealand) of soil pairs displayed in Fig. 3 were calculated (Table 12). Soils on both sunny and shady aspects in New Zealand have the same mean C/N value in Ah horizons and 0–7.5 cm soils (14.9). This is exactly the same mean value as that for Tasmanian A1 horizons under wet forest. In contrast, Tasmanian A1 horizons under dry forest have a mean C/N ratio of 28.5.

4. Discussion

4.1. Soil contrast index

All three SCI groups show lower levels of total C in soils on sunny slopes or under dry forest, compared to levels on shady slopes or under wet forest, confirming that a process or processes favouring organic matter accumulation on shady slopes and under wet forest is/are acting in all soil pairs.

The dominant processes governing the nutrient status of Group 1 soils is weathering and leaching. Cation leaching is more severe on shady slopes than on sunny slopes and this difference, together with the greater weathering prevalent on moister shady slopes, is sufficient in some places to effect an aspect-related change in soil classification (McIntosh, 1986, 1992; McIntosh et al., 2000). In some Mackenzie hill country soils included in this group topsoil erosion has obscured the dominant leaching effect somewhat and caused nutrient loss on sunny aspects—these soils plot to the left within the Group 1 cluster. Most soils show a tendency to have higher C levels on shady aspects (i.e. a negative log C ratio), consistent with the moister conditions and lower average temperatures on these aspects, which favour organic matter accumulation. It can be reasonably concluded that microclimate, through its effects on leaching and organic matter accumulation, is the dominant process controlling natural nutrient levels in Group 1 soils, with erosion playing a significant but minor role in the drier soils of this group.

Group 2 soils straddle the *Y*-axis in Fig. 3, i.e. some soil pairs have nutrient patterns which conform to those expected from microclimate and leaching being the dominant process, whereas some soil pairs (plotting to the left of the *Y*-axis in Fig. 3) have nutrient patterns indicating that topsoil erosion on sunny aspects is over-riding any nutrient-enhancing effect that warmer microclimate and less leaching might have. Field observations of topsoil erosion on sunny aspect soils within SCI Group 2 support this interpretation.

Group 3 soils are confined to Tasmania. They have some properties in common with Groups 1 and 2, for example, C contents of soils are invariably higher in soils on moister sites (under wet forest) than in soils on drier sites (under dry forest) so that their log total C ratio is always negative. However, in contrast to Group 1 soils, which have *higher* levels of nutrients in the drier soil of each pair, Group 3 soils have *lower* nutrients in the drier soil of each soil pair. The drier soils of Group 3 pairs also have texture-contrast profiles where they are developed in quartzose parent materials (i.e. parent materials other than dolerite).[3] The pattern of lower nutrients in the drier of each soil pair in Group 3 soils also applies to P levels, which are unaffected or only slightly affected by aspect differences in Group 1 soils (see Section 3).

Table 13 summarises nutrient differences within Group 3 soils on an area basis. For nutrients other than Mg, Group 3 dry forest sites have 45–58% lower nutrients in soils than wet sites. On the more fertile soils in dolerite the absolute nutrient differences amount to Ca 5.3 t/ha, K 3.0 t/ha, N 5.0 t/ha and P

[3] Texture-contrast soils occur in dolerite, but see a note on their origin in Section 3.2.3.2.

Table 13
Nutrient content of selected soil pairs on an area basis

Measurement	Soils on dolerite			Soils on granodiorite			Soils on granite			Mean % diff.
	Under dry forest (Bream[a] series)	Under wet forest (Wielangta[b] series)	% diff.	Under dry forest (Mckay[c] series)	Under wet forest (Blumont[d] series)	% diff.	Under dry forest (Jensen[e] series)	Under wet forest (Stronach[f] series)	% diff.	
Depth (m)	0.90	0.90		1.00	1.00		0.75	0.75		
Exch. Ca (kg/ha)	3392	8722	−61	1693	3194	−47	314	923	−66	−58
Exch. Mg (kg/ha)	4668	4933	−5	1517	1123	+35	517	416	+24	+18
Exch. K (kg/ha)	1242	4242	−71	1059	1155	−8.3	239	484	−51	−43
Total C (t/ha)	109	159	−31	85	106	−25	65	65	0	−19
Total N (t/ha)	4.8	9.8	−51	4.3	7.7	−44	2.5	4.2	−41	−45
Total P (kg/ha)	674	2335	−71	298	732	−59	753	784	−4	−45

[a] Soil chemistry from McIntosh et al. (2001a); volume weight data from Holloway profile (Laffan et al., 1995).
[b] Soil chemistry from McIntosh et al. (2001b); volume weight data from Excalibur profile (Laffan et al., 1995).
[c] Soil chemistry from Laffan et al. (2002a); volume weight data from Grant et al. (1995a).
[d] Soil chemistry from Laffan et al. (2002b); volume weight data from Grant et al. (1995a).
[e] Soil chemistry from Laffan and McIntosh (2002a); volume weight data from Grant et al. (1995a).
[f] Soil chemistry from McIntosh and Laffan (2005); volume weight data from Grant et al. (1995a).

1.7 t/ha. On the least fertile soils in granite the absolute differences amount to Ca 0.6 t/ha, K 0.25 t/ha, N 1.7 t/ha and P 0.03 t/ha. There is no evidence that either sodicity or salinity has affected profile development in Group 3 soils. Among the cations, Ca is the ''most susceptible to loss by forest management'' (Turner and Lambert, 1986). That Mg balance is an exception to the general trend of other elements (Table 13) may be due to the fact that woody forest vegetation contains much more Ca than Mg (Turner and Lambert, 1986; Adams and Attiwill, 1988) and that, although Ca and Mg are lost in approximately the same proportions that they are present in burnt vegetation (see Section 4.4.2), absolute loss of Ca by fire is greater than loss of Mg. The ratio of Ca and Mg accessions in rainfall (3.8:1) is narrower than the ratio of these elements in standing trees (5.4:1) (Adams and Attiwill, 1988) which could lead to preferential replenishment of soil Mg, as could different Ca and Mg supply rates from weathering and different leaching losses of the two elements, but there is insufficient information on these processes in Tasmania to enable firm conclusions to be drawn.

In attempting to explain why Group 3 soils are so fundamentally different from Group 1 and 2 soils, it could be argued that the Tasmanian Group 3 soils represent extreme cases of processes working in Group 2 soils, i.e. that under dry forest soils must have lost large amounts of nutrients by erosion. But this argument is not supported by observations—all Tasmanian profiles were described on what appeared to be stable sites, and profiles had well-defined horizonation indicating stability and absence of gross disturbance, either biological or physical. We must therefore look for other processes to explain the nutrient depletion in soils under dry forest relative to soils under wet forest in Tasmania.

4.2. C/N ratio

The fact that the mean C/N ratio of A1 horizons in Tasmanian dry forest soils (28.5) is markedly higher than the mean C/N ratio of A1 horizons in Tasmanian wet forest (14.9), and that the mean value of the C/N ratio in A1 horizons in Tasmanian wet forest is identical to the Ah horizon or 0–7.5 cm values for New Zealand sunny aspect soils (14.9) or shady aspect soils (14.9) indicates that the A1 horizons of Tasmanian dry forest soils probably contain a higher proportion of charcoal than their wet forest counterparts. This deduction is supported by the observation by the authors that charcoal is frequently discernible on the soil surface in dry forest, as expected from more frequent forest fires in dry forest (Fig. 2; Duncan and Brown, 1995). We therefore conclude that fires have had a significant effect on A1 horizon carbon chemistry in Tasmanian dry forests, and that other possible effects of fires on Tasmanian soils, particularly on those developed under dry forest, deserve discussion.

4.3. Incidence of fires

In New Zealand the local evidence of paleosols and accompanying charcoal shows that natural fires destroyed areas of forests in the southern South Island in the late Holocene (Molloy et al., 1963) but the intervals between fires were long (McGlone et al., 1995; McGlone, 2001). The exact intervals between forest fires will never be known but are likely to have been centuries or millenia, and of sufficient duration to allow restablishment of podocarp-broadleaf or *Nothofagus* forest.

McGlone et al. (1995) argued that it is only since Polynesian settlers first exploited the southern South Island in about 800 years B.P. that fires have occurred regularly, because the evidence of widespread fires and associated vegetation change after this date is accompanied by a continuous pollen record for bracken (*P. esculentum*), and bracken is a plant adapted to frequent burning. Again, the exact intervals between these relatively recent grassland fires will never be known, but are likely to be decades or centuries.

In Tasmania the influence of fire extends over a much longer time period. Human settlers arrived in Tasmania ca. 34 000 years B.P. (Cosgrove, 1995; Cosgrove et al., 1990). As in New Zealand, the arrival of humans in the forested landscape is highly likely to have increased fire frequency, especially in the drier regions of Tasmania (Jackson, 1968, 2000). In contrast to New Zealand, where forest fires resulted in induced tussock grassland vegetation and subsequent burns were of low intensity (because of the lower biomass of grasslands), fires in Tasmania resulted in regeneration of the fire-adapted eucalypt forests. It is therefore reasonable to assume a history of repeated intense burns in Tasmanian forests. However, these burns will have been much more frequent in the dry forest types, which will have also experienced understorey burns (Duncan and Brown, 1995). Mount (1979) estimated an average fire return interval of 4–20 years in dry forests, and a return interval of 20–100 years in wet forests. Fire intervals may be longer, but where they exceed 100 years the wet eucalypt forest tends to become mixed eucalypt forest, which is defined as eucalypt forest containing rainforest species (e.g. *Nothofagus*) (Duncan, 1996).

Jackson and Bowman (1982) pointed out that ‘‘the incidence of forest fires in Tasmania is virtually dependent on man. Unlike the situation on the Australian mainland, lightning plays but a small part in the incidence of forest fires’’ and they cited records showing that over a 12-year period there were only eight instances of forest fires produced by lightning in Tasmania, with 29 ha of forest burnt. Jackson (1999) developed the argument that the present-day large proportion of disclimax (seral) vegetation in Tasmania developed as a result of increased fire frequency associated with the arrival and influence of man.

It appears possible that after the dry Glacial period Aboriginal fire frequency increased, as dense wet forest encroached on open dry forests in the early Holocene, for botanical evidence from southern Tasmania indicates that forest remnants now surrounded by fire-induced scrub and sedgelands ‘‘may represent a relict dry-sclerophyll community that was once contiguous with [communities in] eastern Tasmania during the height of the last glaciation’’ (Williams and Marsden-Smedley, 2000). In addition, evidence for aboriginal burning for at least 4000 years in northeast Tasmania, and loss of natural fertility and soil podzolisation from about 6500 years B.P. may indicate an attempt by the aboriginal population to keep the encroaching forest out of grasslands formed under the previous drier and cooler climate (Thomas, 1996).

The effect of aboriginal use of fire on Australian mainland woodlands (with an interesting aside on New Zealand forests), was noted by Major Thomas Mitchell in 1848 (quoted by Jones (1969)): ‘‘Fire, grass, kangaroos and human inhabitants seem all dependent on each other for existence in Australia . . . Fire is necessary to burn the grass and form those open forests . . . the native applies that fire to the grass at certain seasons, in order that a young green crop may subsequently spring up and attract and enable him to kill or take the kangaroo with nets . . . But for this simple process, the Australian woods had probably contained as thick a jungle as those of New Zealand or America instead of open forests.’’ The early Tasmanian explorers G.A. Robinson (in 1831) and J.H. Wedge (in 1830) also remarked that the aboriginal population in Tasmania maintained forests free of understorey by regular burning (Duncan, 1996).

4.4. Effects of fires

It appears from the brief summary given above, that fires have been more frequent and probably of greater intensity in Tasmania than in New Zealand, and that in Tasmania fires have been more frequent in dry forests than in wet forests. As the literature on fire effects is large, and sometimes contradictory, it is important to determine what effect these fires might have on soil properties.

4.4.1. 'Hot' and 'cool' burns

Although there is a perception that slash burns and high-biomass burns are hot and prescribed burns and understorey burns are cool, the key difference between these burns may relate more to their duration and ability to burn large woody debris than to temperature. For example, Payton (2002) noted that when a New Zealand tussock grassland containing 27 t/ha of above-ground biomass (including litter) was burnt, temperatures at the soil surface reached 600–700 °C, but that such temperatures were only achieved for a few minutes, and quickly declined after the fire had passed. Control burns in Australian eucalypt forest may reach 320–450 °C and are also short lived (Hatch, 1959). Slash fires and wildfires reported by Ketterings et al. (2000) and Giardina et al. (2000) had a similar or higher temperature range, but such fires may last for hours and large woody debris can burn for days after the main fire front has passed. They therefore have a greater opportunity to transfer nutrients to the atmosphere by convection, and to cause soil effects.

4.4.2. 'Loss' of nutrients to the atmosphere

There have been many studies estimating nutrient losses as a result of burning various forest types. On a clearfelled highly productive wet forest (mixed *Eucalyptus regnans*-rainforest) coupe in the Tasmanian Florentine valley, containing post-harvest slash of 630 t/ha, Harwood and Jackson (1975) found that about half the slash biomass was consumed by a regeneration burn, and calculated that losses to the atmosphere and off the site were: P 10 kg/ha, K 51 kg/ha, Ca 100 kg/ha and Mg 37 kg/ha. As a proportion of the total nutrients contained in the fuel the figures were P 18%, K 17%, Ca 12% and Mg 29%. Assuming the timber harvested from this coupe was approximately double the tonnage of slash left on the ground, one might assume that a hot wildfire would have resulted in greater nutrient losses than calculated above, but such extrapolations from controlled to uncontrolled situations have to be made cautiously, as wildfires may not burn nutrient-rich components (e.g. leaves and twigs) to the same intensity as slash fires or regeneration burns (Raison, 1980), and standing trees may not burn to the same extent as harvest debris on the ground.

On a less productive but ''fertile'' clearfelled site in northeast Victoria containing 270 t/ha of logs and slash (including the original trees, which were not removed), Stewart and Flinn (1985) measured losses of N 50%, P 36%, K 36%, Ca 27% and Mg 29%, i.e. percentage losses of P, K and Ca were about double those of the Tasmanian site, but Mg loss was about the same. When absolute losses were adjusted for ash incorporated into partly burnt soil (0–2 cm), total losses were N 244 k/ha, P 10 kg/ha, K 128 kg/ha, Ca 259 kg/ha and Mg 61 kg/ha. Hopmans et al. (1993) calculated that regeneration burns on a harvested eucalypt area on a texture-contrast soil in granite in Victoria would cause a loss of 268 kg/ha of Ca to the atmosphere, which represented about 25% of the exchangeable Ca in the soil profile. They judged that successive burns would deplete the soil Ca resource.

In dry forests understorey burns result in smaller nutrient losses than described above (e.g. N 74–109 kg/ha; P 2–3 kg/ha; K 12–21 kg/ha; Ca 19–30 kg/ha; Mg 5–10 kg/ha; Raison et al., 1985); however accumulated losses over time may significantly affect ecosystem functioning because dry forests tend to have more fires than wet forests. Waldrop et al. (1987) found that low intensity prescribed litter burns in pine forests of the southeastern coastal plains of the U.S.A. resulted in less Ca in the forest floor and higher concentrations of Ca in the 0–10 cm soil layer, with no net change of the combined Ca content of these layers. It is likely that such burns do not result in significant atmospheric or soil losses of cations and that their main effect is to accelerate nutrient transfer from litter to the mineral soil.

Whether the atmospheric losses quoted above are true losses from the forest ecosystem has been debated. Cook (1994), working on the effects of grassland fires in the continental-scale tropical savannahs of the Northern Territory, Australia, suggested that although ''redeposition [of ash] would be patchy due to wind variations, these transfers should be considered as local redistribution which evens out over time, rather than as losses'', but did not provide data to support this conclusion. An alternative view was put forward by Handreck (1997), who, referring to the work of Ashton (1976) and Bale and Charley (1994) argued ''if such a redistribution is generally in one direction, it would lead to impoverishment of the 'donor' site. Such is apparently the case in hilly areas, where the greater aridity of the north and north-western slopes [in the southern hemisphere] allows more frequent fires and transfer of P to other slopes.''

While not denying that a proportion of particulate ash in smoke from forest fires will reach the ground nearby, it seems likely that in the hilly and steep terrain of eastern Tasmania, bordered in the east in pre-European times by the ''sparsely wooded plains'' of the Tasmanian Midlands basin (Fensham and Kirkpatrick, 1989) and in the far south by Storm Bay, and to the west by the Tasman Sea, the 'impoverishment' model of Handreck (1997) rather than the 'continental redistribution' or 'no net loss' model of Cook (1994) will apply. We suggest that there will be a net eastward drift of nutrients from forest fire smoke due to the prevailing westerly and northwesterly winds, with loss of nutrients over the ocean. Furthermore there will be a net nutrient loss from more frequently burnt drier windward slopes to less frequently burnt moister leeward slopes.

4.4.3. Loss of nutrients by erosion

Burnt topsoils and surface ash are prone to erosion by runoff (Walker et al., 1986; Raison et al., 1990; Saá et al., 1994; Thomas et al., 1999; Wallbrink et al., 2005) and wind (Giardina et al., 2000). Losses of nutrients (N, P and K) from ash and soil in runoff following bushfires in Portugal were 3–4 orders of magnitude greater than those on unburnt sites (Thomas et al., 1999), and continued for at least 3 years after the fire. In dry coastal forests (Mackay and Robinson, 1987) particulate matter losses were found to be small, amounting to 1–4 kg/ha per year, and decreasing to zero after 3 years. Raison et al. (1985) considered that any loss of particulate ash into streams is likely to be significant because of the concentration of nutrients, particularly cations, in ash; these authors recorded that black ash contained about 10 times the concentrations of P, K, Ca and Mg found in litter, and grey ash, resulting from more complete combustion, contained 10–30 times the concentrations of P, K and Mg and 50 times the concentration of Ca found in litter. Erosion losses are likely to be greatly influenced by local factors such as slope (Saá et al., 1994), occurrence of high-intensity rainfall (Hudson et al., 1983; Wallbrink et al., 2005), speed of re-establishment of vegetation cover, soil texture, structure, and water repellency (De Bano, 2000; Shakesby et al., 2003). Wallbrink et al. (2005) investigated erosion of ash and soil after the summer bushfires of 2001/2002 in the hilly Lake Burragorang catchment, Sydney by ^{7}Be, ^{210}Pb and ^{137}Cs techniques. The fires burnt both the understorey and forest canopy, and they noted large amounts of ash and charred clay–organic matter aggregates reaching streams. Ash movement was accelerated by six 50–100 mm rainfall events within 2 months of the fires. From isotope work they calculated that 28–35% of the burnt organic matter forming the ash layer had been lost from major landscape units (Wallbrink et al., 2005).

4.4.4. Returns of nutrients in rainfall

Adams and Attiwill (1988) measured rainfall nutrient additions in northern Tasmania and under mean annual rainfall of 790–2010 mm recorded annual additions of N 4–8 kg/ha and P 0.2–0.8 kg/ha, and mean values of K 3.5 kg/ha, Ca 11 kg/ha and Mg 3 kg/ha. Returns in rainfall over an 80-year rotation (a typical rotation length for managed Tasmanian eucalypt forests) in northern Tasmania would therefore be N 320–640 kg/ha, P 16–64 kg/ha, K 28 kg/ha, Ca 88 kg/ha and Mg 24 kg/ha. When these rainfall nutrient inputs are compared to losses to the atmosphere reported above (assuming no returns to the ground), a simple conclusion would be that rainfall could replenish N and P losses from a 1 in 80-year fire that is hot enough to cause substantial losses by ash convection, but rainfall would not replenish K, Ca and Mg losses. However, on a burnt wet forest site in southern Tasmania, Ellis and Graley (1983) concluded that ''nutrients lost from the area as particulate ash are in quantities that will probably be replaced in rainfall within 15–20 years''.

Although simple calculations might indicate that long-term atmospheric additions of nutrients are ample to make up for losses resulting from fires (as proposed by Ellis and Graley (1983)), there appear to be mechanisms operating that prevent these additions from being effective in restoring nutrients lost. This deduction can be readily demonstrated by comparing the nutrient additions since soils have been stable at the end of the Last Glacial with present soil nutrients. For example, the P accession under 1000 mm rainfall in Tasmania is approximately 0.5 kg/ha/yr (Adams and Attiwill, 1988) which amounts to 5 t/ha over 10 000 years. This figure far exceeds the total P difference (1.7 t/ha) between Bream and Wielangta soils to 90 cm depth (Table 12), and the total P in either soil. So even on the clay-rich and moderately P-retentive Bream soil steady P losses must occur, otherwise rainfall would have restored P levels to about the values found in the nearby Wielangta soils formed on the same parent rock and parent material. In sandy soils the discrepancy between total soil P and other nutrients and long-term rainfall accessions is greater, and it must be assumed that either the soils are incapable of retaining added nutrients, or that regular losses occur from the surface soil and vegetation, or that both processes are operating.

4.4.5. Changes in nutrient availability

Many studies have shown that forest burning increases the availability of nutrients in surface soils (e.g. Humphreys and Lambert, 1965; Waldrop et al., 1987; Raison et al., 1990, 1993; Binkley et al., 1992; Chambers and Attiwill, 1994; Ludwig et al., 1998) and is associated with a short-term rise in topsoil pH which may be as large as 2.5 pH units (Ellis and Graley, 1983; Guinto et al., 2001). However, initial pH rises may be reversed later (Binkley et al., 1992). While short-term increases in nutrients have generally been regarded as positive, some authors have cautioned that negative long-term effects may occur. For example Wells (1971) noted that the short-term nutrient increase resulting from fire may in time, become a nutrient deficit, resulting in lower rates of growth on burnt sites than on unburnt sites. Kellman et al. (1985), in a lysimeter study following a savannah fire in Belize, measured significant increases in solution Mg, K and Na shortly after fires and suggested that “frequently repeated fires may elicit longer-term changes in soil fertility if the losses are incremental”, while acknowledging that the soils they studied were resistant to “acute leaching losses”.

We suggest that such losses are most likely if the cation exchange capacity of the soil is exceeded by large accessions of nutrients in readily soluble form at the soil surface, although nitrate-driven cation leaching could also be important. Small incremental losses (if they occur) are of course more difficult to record and to assess in a statistically meaningful manner than large increases on the soil surface, given the large natural variation in soil properties (Binkley et al., 1992). These authors found that pH of both the 0–10 cm and 10–20 cm layer of soils in pine forest subjected to a prescribed burn declined by 0.3 pH units 4 years later. Mean values of exchangeable Ca and Mg in 10–20 cm soils indicated that levels of these cations may have decreased 4 years after the burn (which would, in part, explain the pH decline), but levels were not significantly different ($P > 0.1$) from those in control plots. The study by Binkley et al. (1992) indicates that caution should be applied before making generalisations from topsoil (e.g. the 0–10 cm soil changes recorded by Waldrop et al. (1987)). Although Ca and Mg may become more available after intense slash fires in clearfelled eucalypt areas, over time (4 years) less soluble Ca and Mg salts may form in surface soils (Ludwig et al., 1998).

4.4.6. Other chemical changes

Fires in which surface temperatures exceed 200 °C can decrease clay content of soils and alter soil mineralogy (Ketterings et al., 2000; Sertsu and Sánchez, 1978). Although such temperatures cannot be reached until all the soil moisture has been evaporated, and modelling indicates that temperatures fall rapidly with increasing soil depth and at 10 cm depth seldom rise above 100 °C (Preisler et al., 2000), Ketterings et al. (2000) pointed out that where stumps are present burning can extend 50 cm deep into the soil, and such burnt holes, similar to those photographed on the Australian mainland (Fig. 4), are common in burnt eucalypt forests of Tasmania. Protection from burning resulted in large increases (40–50%) of soil C in the 0–5 cm layer of savannah soils of Zimbabwe (Bird et al., 2000), but an intense slash fire in a eucalypt forest resulted in an increase of soil C (Ludwig et al., 1998), which was attributed to the contribution of dead roots to the <2 mm analysed soil fraction. It may not be possible to generalise about different sites: 40 years of burning on a dry sclerophyll site in Queensland, Australia did not lead to any loss of C and N in topsoil (0–10 cm) initially containing 2.2% C, but on a wet sclerophyll site in which the topsoil

Fig. 4. Eucalypt tree stump burnt to >30 cm depth in the Canberra wildfires of March 2003, photographed in March 2004. The soil is a texture-contrast soil formed in granite, on a slope of 15°. Note that rain after the fire has washed ash from the soil surface, leaving the white sandy upper horizons of the texture-contrast soil exposed. Notebook is 15 cm across.

contained 4.3% C biennial burns over 22 years resulted in lower C and N levels (Guinto et al., 2001).

Weathering releases small but steady amounts of P and cations (Velbel, 1985). In soils on glacial tills in New Hampshire, U.S.A., cation release into streams by weathering was Ca 5.0 kg/ha, Mg 1.8 kg/ha and K 0.7 kg/ha (Likens et al., 1967). A lysimeter study by Knight (1987) in rhyolitic tephra terrain in New Zealand produced similar figures: Ca 6.0 kg/ha, Mg 2 kg/ha and K 4 kg/ha.

Not all nutrient release by weathering may be retained by the soil, or be available for uptake by plants, particularly in subsoils where roots may be uncommon: Hopmans et al. (1993), quoting Marchand (1971), estimated that weathering from granite in Victoria could supply Ca 17 kg/ha/yr, but that only 7 kg/ha will be retained in the soil and only a proportion of this retained Ca would be available for uptake to replenish surface losses.

The contribution to soil N of N fixation after fires is beyond the scope of this study. N fixation rates may vary between forest types, sometimes appearing to be sufficient to replace burning-induced losses (Adams and Attiwill, 1984) and sometimes not (Hamilton et al., 1991). In other studies it has been suggested that N inputs approximately balance losses from burning (e.g. Hopmans et al., 1993).

4.4.7. Summary of fire effects

It is evident from the above discussion that fires can lead to direct losses of nutrients to the atmosphere, and that the hotter and more prolonged the fire, the more likely losses will occur. In continental land masses returns of particulate ash may balance losses, but there is little information on this subject. In hilly country and country bordered by non-forested areas, or sea, atmospheric losses are unlikely to be balanced by gains from ash fall. There have been studies which show that nutrient losses also occur by erosion of ash. Erosion is a particularly effective method of removing nutrients from a burnt site because the concentrations of nutrients in ash far exceed concentrations in soils and litter, and ash is readily removed by runoff. Loss of vegetation cover and increased soil water repellency after fires contribute to increased runoff and surface erosion after fires. There is only limited evidence to suggest enhanced leaching of nutrients after fires, although intuition suggests that after fires, a combination of high levels of nutrients (some water-soluble) at the soil surface, and death of the roots associated with understorey (and in some cases canopy) vegetation will increase the risk of leaching, particularly if heavy rainfall occurs before regenerating trees, shrubs and grasses have established. Although, in theory, accessions of nutrients in rainfall should balance losses from fires, in practice this does not seem to be the case.

4.5. Feedback mechanisms

It is evident that in order to decide whether burning produces net nutrient change in soils, using the nutrient balance principle *in isolation*, would require detailed quantification of many processes that are extremely difficult to measure (e.g. long-term leaching losses, erosion, weathering inputs) and costly to quantify, particularly if the aim is to establish long-term trends. An alternative to the modelling approach is to consider the empirical evidence of vegetation and soil profile change, which can be assumed to integrate the effects of numerous processes acting over the history of a site.

As early as 1981 Bowman and Jackson (1981, p. 359) argued that "the effect of fire is to reverse the accumulation of nutrients in successional vegetation" and later (Bowman et al., 1986; Jackson, 2000), building on the observations of Boerner (1982), they argued that the effect of fire is a *feedback mechanism*: once rainforest or wet eucalypt forest has been burnt, the vegetation occupying the site becomes more prone and progressively adapted to fire—the burnt site is more likely to support another fire as fire frequency increases, and ‘‘recurrent fire encourages vegetation with lower nutrient requirements’’. Jackson and co-authors developed this feedback mechanism theory to explain the development of low-altitude sedgelands on nutrient-poor parent materials in southwest Tasmania (Bowman et al., 1986; Jackson, 2000). While Jackson and co-workers did not attempt to explain different soil morphological development processes in forested land, their analysis was groundbreaking in that they considered the soil to be a dynamic and integral part of an evolving ecological system, not just a nutrient bank with a net positive or negative balance after fires.

Building on the arguments of Jackson and co-workers, we can conclude from the studies reported above that although fires may increase the short-term nutrients at the soil surface they also increase the risk of long-term loss of nutrients from standing vegetation and litter. Atmospheric losses, losses by erosion and runoff and losses by leaching are likely to occur after fires. Losses by leaching are the least well documented. We can be certain, however, that in contrast to the steady nutrient returns to the mineral soil that occur as litter decomposes in a wet eucalypt forest or in a forest not subjected to fire, that the soil in a forest subjected to fires receives most of its nutrients in pulses, and the nutrient contribution pulses occur when the vegetation is least able to absorb nutrients (because of death of roots) and when infiltration and runoff is higher than under unburnt forest (because of decreased evapotranspiration; Vertessy et al., 2001). Nutrient loss is likely after fires, until vegetation re-establishes, root activity is restored and evapotranspiration returns to normal levels. Therefore, paradoxically, the increased availability of nutrients at the soil surface after fires leads to decreased total nutrients and decreased nutrient availability in the long term.

The net effect for a site subjected to frequent fires must be that the vegetation has to adjust to lower fertility: not only will some nutrients have been lost to the atmosphere, but those that are returned in a pulse to the soil surface are unlikely to be fully utilised and are prone to further losses, and the steady nutrient return to the soil from litter breakdown that occurs in unburnt forests will not be available. To compensate for the absence of steady nutrient returns trees are likely to source nutrients from deep soil layers, as described for Ca by Khanna and Raison (1986). Such nutrients, in time, will either be retained in vegetation or returned to the soil surface as litter, where they will also be prone to loss after fires.

We therefore propose that an ecosystem subject to frequent fires is potentially more ‘leaky’ of nutrients than a forest ecosystem in which fires are rare, because the latter ecosystem not only has fewer direct losses of nutrients (in the form of smoke particles and erosion) but also is less exposed to repeated ‘windows of risks’ for loss of nutrients when heavy rainfall follows fires.

The above nutrient models for dry and wet forests are supported by soil pH trends under dry and wet forest, respectively. In Kurosols under dry forest (Tables 8–10) the pH trend is that of *lower* values near the soil surface than in deeper soil horizons, which reflects the long-term *depletion* of cations from the surface horizons of these soils. In contrast, in Dermosols under wet forest the trend of soil pH is for values to be *higher* near the soil surface (Tables 8–10) and to decrease with increasing depth, which reflects net *accumulation* of soil nutrients at the soil surface, resulting from nutrient cycling by trees over hundreds (and in some cases, probably thousands) of years. Consequently, we argue, that from similar soil beginnings, a fire-prone ecosystem proceeds irreversibly down a pathway of incremental nutrient loss and increasing susceptibility to further fires, because the decreasing nutrient status of the ecosystem encourages fire tolerant forest communities, i.e. the feedback mechanism proposed by Bowman et al. (1986) and Jackson (2000) operates in dry forests as well as in the sedgelands studied by these authors.

That low-nutrient soils are particularly prone to nutrient depletion is shown not only by Jackson's (2000) examples of forests developed on quartzitic soils being converted over time to sedgelands, but by a study of Sands (1983) that demonstrated that burning a nutrient-poor sandy podzol resulted in large reductions in organic matter, total N, available P and CEC which were detectable 24 years after the fire. An important consideration is that such soils, with their depleted vegetation (relative to infrequently burnt or unburnt sites) will also be less capable than nutrient-rich soils of retaining added nutrients, whether these nutrients are derived from rainfall accessions or from nutrient cycling. In extreme cases of texture-contrast soils (Jensen and Mckay soils), A1 and A2 horizons are sandy and have P retention values of 0–2% (Table 8). In such horizons there will be virtually zero physical adsorption of P; the small amounts of total P in these horizons is probably entirely in the organic form.

4.6. Soil morphology and soil genesis

While repeated direct and indirect depletion of nutrients by fires can explain the nutrient differences of soils under dry and wet forests in Tasmania, the profile differences between these soils also require explanation, particularly the ubiquitous presence of sandy A1 and A2 horizons and texture-contrast soils under dry forest on quartzofelspathic parent materials (Fig. 2) and the contrasting uniform- or gradational-texture soils under wet forest (Appendix 1).

Three processes favour the persistence of gradational soils under wet forest. These are: (1) greater nutrient cycling and organic matter cycling and carbon accumulation under the greater biomass of wet forests compared to dry forests; (2) more soil mixing by soil fauna (particularly earthworms) in the moister wet forest soils (Laffan and Kingston, 1997) compared to dry forest soils; and (3) stabilisation of clay by humus–clay–cation linkages (Muneer and Oades, 1989; Percival et al., 2000; Clough and Skjemstad, 2000; Ahmed et al., 2002a,b) which is likely to be more significant in the more organic matter rich and higher quality (lower C/N ratio; Table 12) organic matter of gradational soils under wet forest. Conversely, destruction of stable organic matter–clay linkages, changes of clay mineralogy to relatively inert clays, and coarsening of soil texture after hot fires (Ketterings et al., 2000; Sertsu and Sánchez, 1978), together with litter and organic matter destruction that creates an inimical soil environment for earthworms and mixing fauna (Laffan and Kingston, 1997) is likely to favour clay eluviation and formation of texture-contrast soils. We point out that repeated heating of the mineral soil, even if it is only to 1 mm depth in any one fire, will cumulatively tend to destroy clay–organic matter linkages, and local burning and heating to greater depth undoubtedly occurs (Fig. 4; Ketterings et al., 2000). We therefore conclude that under the same rainfall, soils subject to frequent fires will be more prone to disassociation of clays from their organic matter "glue" than soils subject to infrequent fires, and will therefore be more prone to clay eluviation.

Each Tasmanian soil pair described in the case studies and each nutrient pair in Fig. 3 is formed under a very similar climate, and the soils of each pair differ not only in their vegetation cover and nutrient status, but also in their profile form and soil classification. Except for the described soils in dolerite, the soils under dry forest are texture-contrast soils (Kurosols) and have strong evidence of inter-horizon clay eluviation, whereas soils under wet forest are uniform or gradational soils (Dermosols) with only weakly expressed or no clay eluviation features. This evidence, plus the New Zealand evidence of a relatively short history of fires, the predominance of low-biomass tussock grassland fires in New Zealand, and absence of texture-contrast soils in that country leads us to deduce that in Tasmania texture-contrast soils are typically formed on sites subject to frequent fires, and have in fact been formed by the effects of repeated fires.

The occasional association of gradational or uniform soils with dry forest (rather than the more usual wet forest cover) requires some explanation. There appear to be three reasons why this association occurs: (1) the soil parent material is nutrient- and clay-rich and therefore less prone to clay eluviation (for example, Holloway and Bream soils in dolerite; Tables 11 and 12); (2) a relatively short history of fire—enough to induce dry forest but not enough to induce texture-contrast profiles (see for example the Wurrawa profile under dry forest; Laffan and McIntosh, 2002b); and (3) bioturbation by earthworms in moister soils having impeded drainage (Laffan and Kingston, 1997). The converse association of wet forest with texture-contrast soils also sometimes occurs (e.g. Paris soils described by Grant et al. (1995a)); this association chiefly occurs in higher rainfall areas (>1200 mm mean annual rainfall) and the vegetation type it supports is not tall wet forest with *E. regnans* (Fig. 2B) but typically *E. obliqua* with a mixed bracken/heath/broadleaf understorey. We conclude that this intermediate vegetation type develops on texture-contrast soils which have a long fire history but which have not experienced a fire for about 100–200 years.

Since fire is not a common natural occurrence in Tasmania (because lightning is usually accompanied by rain, in contrast to lightning storms on the mainland of Australia; Jackson and Bowman, 1982) we can postulate with some confidence that the texture-contrast soils of Tasmania are not only fire-induced, but are in part *anthropogenic*, i.e. they have been produced by the effects of repeated human-lit fires since about 10 000 years ago when soil parent materials stabilised, or over a longer period on stable sites.

In contrast the New Zealand forested landscape was only subjected to frequent fires after 800 years B.P, when humans arrived (McGlone, 2001). Moreover, in New Zealand there were few fire-adapted trees and shrubs (and none of significant stature and biomass comparable to the *Eucalyptus* genus) able to invade the burnt landscape—instead the vegetation was invaded by tall tussock (*Chionochloa*) and short tussock (mostly *Poa* and *Festuca*) grasslands lacking the biomass and therefore the nutrient loss potential of Tasmanian forests. So after the first fires that destroyed the indigenous forests in New Zealand, further fires would have had neither the heating or nutrient removal effects of fires in Tasmania. We therefore conclude that there has been insufficient time, organic matter destruction or nutrient removal in New Zealand forest soils to allow formation of texture-contrast profiles.

5. Summary

1. The dominant soil patterns in forested or previously forested landscapes in southern New Zealand and Tasmania are described by comparing properties of soil pairs on adjacent sunny and shady aspects in hill country of the South Island of New Zealand and by comparing soil properties of soil pairs under adjacent 'dry' and 'wet' eucalypt forest under the same macroclimate in Tasmania. This comparison broadly follows the paired site retrospective approach recommended by Raison et al. (1993).
2. The expression of soil pair differences as a ratio of log total C values for each pair divided by log exchangeable Ca + Mg + K ratio for each pair (here named the soil contrast index or SCI) allows comparison of soils on parent materials of different absolute nutrient contents. SCI plots show that the soil pairs can be split into three groups. Almost all soil pairs show a trend of having higher total C content on moister sites (i.e. on shady slopes in New Zealand and under wet eucalypt forest in Tasmania). SCI Group 1 soil pairs are stable New Zealand soils in which exchangeable Ca + Mg + K values are higher on drier sunny slopes than on moister shady slopes. SCI Group 2 soil pairs are New Zealand soils in which soils on sunny slopes display evidence of topsoil erosion by wind; consequently some Group 2 soil pairs have soils on dry (sunny) aspects with lower levels of exchangeable Ca + Mg + K than soils on moister (shady) aspects. SCI Group 3 soil pairs are Tasmanian. Soils on drier sites (under dry forest) invariably have lower exchangeable Ca + Mg + K values than soils on moister sites (under wet forest), which is the reverse of the pattern in SCI Group 1 soils in New Zealand.
3. The mean C/N ratio of A1 horizons in Tasmanian dry forest (29) is markedly higher than the mean C/N ratio of A1 horizons in Tasmanian wet forest (15), or values for New Zealand sunny aspect soils (15) or shady aspect soils (15), indicating a higher proportion of charcoal in Tasmanian dry forest soil A1 horizons.
4. Under dry forest in Tasmania soils in quartzofelspathic parent material (derived from sandstone, granite and granodiorite) are generally texture-contrast Kurosols, whereas soils under wet forest are generally Dermosols with uniform or gradational texture profiles. The texture-contrast soils display strong clay eluviation: upper horizons have sand or sandy loam textures and lower horizons have clayey textures. In contrast, texture-contrast soils are all but absent in New Zealand, and do not occur in the previously forested areas described in this paper.
5. For the profiles for which volume weight can be calculated, the wet forest minus dry forest soil nutrient differences amount to a mean 'loss' of 58% of the exchangeable Ca and 43% of the exchangeable K under dry forest. Total N and total P are lower by 45% in soils under dry forest, compared to soils under wet forest. There is no evidence to indicate that the lower values of exchangeable Ca and K, or of total N and P under dry forest, are caused by erosion of mineral soils.
6. We propose that the process causing both (1) the marked difference of nutrients in soils of different

moisture status in Tasmania; and (2) the development of texture-contrast soils in Tasmania, is fire. Fire depletes nutrients in forests by causing direct losses to the atmosphere, by runoff, and by leaching. Frequent fire encourages vegetation that is not only fire-tolerant, but vegetation that is adapted to nutrient-depleted less fertile sites. Thus nutrient loss by fire encourages further fires and nutrient loss, and frequent fire is a feedback mechanism causing progressive soil nutrient depletion over a period of several thousand years. Fire also destroys litter and soil organic matter and will have a negative effect on soil-mixing fauna (e.g. earthworms), soil homogenisation and the ability of organic matter to bind with clay. Consequently fire will encourage clay eluviation.

7. In New Zealand natural fires have been infrequent, and frequent fires have only occurred since ca. 800 years B.P when the first humans arrived, so fire has had little effect on soil development. Furthermore, the absence of fire-responsive trees like the eucalypt genus in the New Zealand flora has meant that the forest destroyed by natural or man-induced fires has been replaced by low-biomass tussock grasslands rather than by trees. Consequently the potential for large and repeated nutrient losses from these New Zealand grassland ecosystems has been much less than the potential for nutrient loss from Tasmanian regrowth forests. In Tasmania natural fires are also infrequent, but fire frequency is likely to have increased markedly after the arrival of the aboriginal population ca. 34 000 years ago. We conclude that during the Holocene dry forest types in Tasmania have extended in area, and that the feedback mechanism of nutrient loss has encouraged the development of texture-contrast soil profiles in areas that would otherwise have retained uniform- or gradational-texture soils.
8. A corollary of the above conclusion is that nutrient-depleted texture-contrast soils in Tasmania are in part anthropogenic—they have formed since the arrival of human settlers in Tasmania 34 000 years B.P., and their development probably dates from ca.10 000 years B.P., when soils parent materials stabilised.

Acknowledgements

The authors thank Judy Sprent, librarian at CSIRO, Hobart, and Andrew Wilson, librarian at Forestry Tasmania, Hobart, for literature searches and Trevor Webb and Ian Lynn, Landcare Research, Lincoln, Canterbury, for providing New Zealand soil information. The constructive comments of Dr. Bill Lee and two anonymous referees for comments on an earlier version of this paper are greatly appreciated.

Appendix 1. Soil classification

Detailed soil classification (in bold) is given according to the system used in the country of origin. Approximate classification is given for other systems. The New Zealand Soil Classification (Hewitt, 1998) is referred to as the NZSC, the Australian Soil Classification (Isbell, 1996) as the ASC, and USDA Soil Taxonomy (USDA, 1998) is referred to as USDA ST.

Soil pair	New Zealand	
	On shady slopes	On sunny slopes
NZ1		
Soil name	Kaiwera	Kaihiku
NZSC	**Acidic** or **Typic Orthic Brown**	**Argillic Orthic Melanic**
ASC	Dermosol	Dermosol
USDA ST	Dystrudept	Eutrustept
NZ2		
Soil name	Venlaw	Kaiwera
NZSC	**Acidic Allophanic Brown**	**Acidic** or **Typic Orthic Brown**
ASC	Dermosol	Dermosol
USDA ST	Cryudept	Dystrudept

Appendix1. (*Continued*)

Soil pair	New Zealand	
	On shady slopes	On sunny slopes
NZ3		
Soil name	Tekoa "shady"	Tekoa "sunny"
NZSC	**Allophanic Brown**	**Allophanic Brown**
ASC	Dermosol	Dermosol
USDA ST	Dystrudept or Cryudept	Dystrudept
NZ4		
Soil name	n.a.	n.a.
NZSC	**Brown**	**Pallic**, **Recent** and **Brown**
ASC	Kandosol and Tenosol	Kandosol and Tenosol
USDA ST	Dystrudepts	Haplustepts
NZ5		
Soil name	n.a.	n.a.
NZSC	**Pallic** and **Brown**	**Pallic**, **Recent** and **Brown**
ASC	Kandosol and Tenosol	Kandosol and Tenosol
USDA ST	Dystrudepts and Haplustepts	Haplustepts and dystrudepts
NZ6		
Soil name	Not named	Not named
NZSC	**Typic Immature Pallic**	**Typic Orthic recent**
ASC	Tenosol	Tenosol
USDA ST	Dystrudept	Haplustept

Soil pair	Tasmania	
	Under wet forest	Under dry forest
Tas1		
Soil name	Maweena	Retreat
NZSC	Ultic	Brown
ASC	**Acidic-Mottled Mesotrophic Brown Dermosol**	**Bleached Dystrophic Brown Kurosol**
USDA ST	Hapludult	Haplustult
Tas2		
Soil name	Stronach	Jensen
NZSC	Brown	Ultic
ASC	**Acidic Mesotrophic Brown Dermosol**	**Bleached-Vertic Mesotrophic Yellow Kurosol**
USDA ST	Dystrudept	Haplustult
Tas3		
Soil name	Blumont	Mckay
NZSC	Brown	Ultic
ASC	**Acidic Mesotrophic Brown Dermosol**	**Bleached-Mottled Mesotrophic Yellow Kurosol**
USDA ST	Dystrudept	Haplustult
Tas4		
Soil name	Excalibur	Holloway
NZSC	Firm Brown or Orthic Brown	Firm Brown or Orthic Brown
ASC	**Red Eutrophic Ferrosol**	**Red Mesotrophic Ferrosol**
USDA ST	Eutrudept	Dystrudept
Tas5		
Soil name	Wielangta	Bream
NZSC	Firm Brown or Orthic Brown	Firm Brown or Orthic Brown
ASC	**Red Eutrophic Ferrosol**	**Red Mesotrophic Ferrosol**
USDA ST	Eutrudept	Dystrudept

n.a.: not applicable.

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