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Rapid deforestation of South Island, New Zealand, by early Polynesian fires

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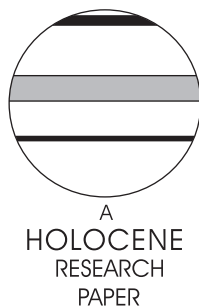
Abstract: In most parts of the world where people have colonized and modified their landscapes for several millennia or more, it is often difficult to discriminate anthropogenic burning from natural fire regimes that are linked to climate regimes. New Zealand provides a unique setting for identifying human influence on fire occurrence because it was settled recently (*c*. AD 1280) at a time when climates are considered to be similar to today. Late-Holocene pollen and charcoal records from New Zealand provide striking evidence for initial Polynesian (Māori) arrival being strongly associated with widespread burning and loss of native forest. The duration of initial forest clearance and the spatial pattern of burning that led to this transformation are still poorly understood. We present high-resolution charcoal and pollen analyses of sediment cores from five lakes, located on the deforested eastern side of the Southern Alps. These records document the local fire history of the last 1000 years and the response of vegetation and watersheds to burning. Our results suggest that one to several high-severity fires occurred within a few decades of initial Māori arrival, and this 'Initial Burning Period' (IBP) resulted in the majority of forest loss and erosion. Changes in sedimentation rates, soil chemistry and magnetic susceptibility occurred simultaneously with the first fires at some sites, and marked the end of the IBP at others, suggesting substantial and rapid alteration of watershed vegetation, soil and biochemistry. Timing of the beginning of the IBP varied across sites but the duration of this period was brief (decades to a century). Our results suggest that Māori burning of native forests was deliberate and systematic. These forests had no previous history of fire and thus showed little resilience to the introduction of a new disturbance.

Key words: Fire history, deforestation, New Zealand, Māori, Polynesians, charcoal records, *Nothofagus*, podocarp, bracken.

Introduction

The extent to which indigenous peoples modified their environment is keenly debated, and this discourse places our perception of present landscapes along a spectrum that ranges from near-pristine to mostly anthropogenic creations (eg, Sauer, 1950; Cronon, 1996; Vale, 1998; Mitchell, 2005). It is argued that hunter-gatherers and early pastoralists in many parts of the world created fire-adapted ecosystems. In Australia, for example, the concept of fire-stick farming (Jones, 1969) attributes extensive landscape modification to deliberate and frequent burning by Aborigines (Nicholson, 1981; Bowman, 1998). Likewise, in North America, it has been suggested that Native American-set fires were neither accidental nor small but rather large-scale efforts at forest management (Denevan, 1992). Conventional wisdom in Europe and Africa also asserts the liberal use of fire by prehistoric peoples (Pyne, 1995, 1997; Burchard, 1998;

Bird and Cali, 1998). The contrary view is that natural factors, particularly climate and vegetation type, are and have been more important than humans in determining fire regimes. Palaeoecological results from these same continents have been interpreted as evidence for indigenous people using fire at the local scale, but having a more variable impact on vegetation at the regional scale (Kaye and Swetnam, 1999; Whitlock et al., 2003; Black and Mooney, 2006; Bird *et al*., 2008). New Zealand is fundamentally different to the situation in Australia and North America because the arrival of humans was relatively recent and subsequent changes in climate have been minor (Lorrey *et al*., 2008; Wilmshurst *et al*., 2008).

Climate change and anthropogenic burning are synergistic processes on most continents, with climate change directly altering vegetation and fuel conditions, and indirectly affecting resources that determine human lifeways as well the sensitivity of vegetation to human-induced disturbance. For example, in Australia, where the biota has evolved with fire over many millennia, it is

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difficult to separate the effects of climate change and anthropogenic activities on past fire regimes (Mooney *et al*., 2007). In New Zealand, however, the arrival of Polynesians (Māori) at *c*. AD 1280 (Wilmshurst *et al*., 2008) had a profound influence on New Zealand ecosystems. Palaeoecological records show that both islands of New Zealand were extensively forested before human settlement (Cumberland, 1962; McGlone, 1983, 2001; Mark and McLennan, 2005), but by the time of European arrival in the mid-nineteenth century, up to 40% of the forest had been lost (McGlone, 1989). The age of the earliest dated archaeological sites is associated with numerous faunal extinctions, evidence of the introduced Pacific rat (*Rattus exulans*) and evidence of widespread deforestation (Wilmshurst *et al*., 2008). Unprecedented levels of charcoal particles in palaeoecological records dating to this time point to anthropogenic burning as the cause of forest loss (McGlone, 1989, 2001; McGlone and Wilmshurst, 1999; Ogden *et al*., 1998; Newnham *et al*., 1998). Although over 150 pollen records from New Zealand show this association between fire, arrival of people and vegetation change, questions still remain with respect to the timing, pattern and consequences of Māori burning.

A suite of new charcoal analysis techniques (Whitlock and Larsen, 2001; Whitlock *et al*., 2006) now permits better characterization of past fire regimes than the conventional approaches previously applied in New Zealand to detect past fires (eg, McGlone and Wilmshurst, 1999). In particular, high-resolution macroscopic charcoal analysis of contiguously sampled lake-sediment cores allows reconstruction of local fires and enables individual fire episodes to be discerned, often with decadal precision. For example, fire-regime characteristics, including the frequency, size or intensity, and severity (eg, crown or surface fires) of past fire episodes can be inferred from aspects of the charcoal data, and identification of charcoal particles allows the source of fuel to be identified to grasses, ferns, hardwoods or conifers (Enache and Cumming, 2006; Whitlock *et al*., 2006). These new approaches, in combination with pollen-based vegetation reconstructions, provide a better understanding of the fire-regime shifts and associated changes in vegetation.

In this paper, we apply new charcoal analysis techniques with standard palynological and lithological analyses to sediment cores collected from five lakes in the deforested south-central South Island of New Zealand to address the following questions: (1) when did the Initial Burning Period (IBP) occur in the southern South Island, was it synchronous, and how did it compare with the age of the event elsewhere in New Zealand? (2) Was the IBP accomplished by a single fire or a sustained period of burning? (3) How quickly did local watersheds and vegetation respond to the IBP, and what was the nature of the response? (4) How did the initial ecological transformation compare with later European land-use activities? (5) Was climate change an important factor influencing the timing of the IBP or the occurrence of fires during the Māori and European periods? Five small lakes with minimal stream inflow and small catchments were selected within three largely grassland and fern-shrubland covered landscapes: Wanaka, Wakatipu and Te Anau (Figure 1). Although existing pollen records from these areas reveal that they were deforested by fire at *c*. 700–800 years ago (McGlone and Wilmshurst, 1999; Wilmshurst *et al*., 2002), paradoxically, the dearth of archaeological evidence from these areas suggests only small and transient Māori populations used the area prehistorically.

Study area

The study regions lie in the southeastern hill country and mountains of the southern South Island of New Zealand, an area which ranges between 250 and 1900 m elevation (Figure 1). The climate is cool year-round and wet with most of the precipitation occurring in the winter months (Table 1). The eastern side of the Southern Alps is
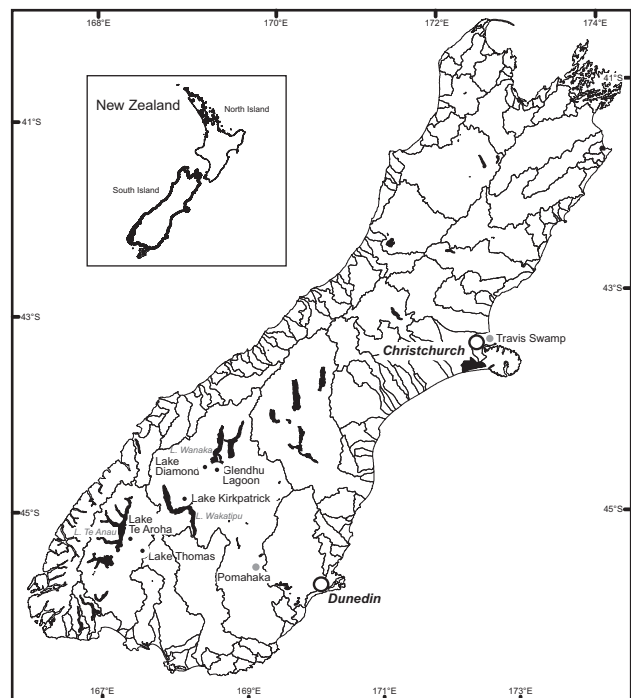

Figure 1 Location of study sites (black circles) and other sites (grey circles) referred to in text with catchments (black outline) and major lakes (black shading) shown, South Island, New Zealand

characterized by steep rainfall and elevational gradients. The original extent and composition of the forest are inferred from pollen records, isolated surviving forest stands and remnant wood (McGlone *et al*., 1995; Wardle, 2001). Pollen, wood and charcoal records indicate that closed forests of mountain totara (*Podocarpus hallii*) and mountain toatoa (*Phyllocladus alpinus*) were once widely dominant east of Lake Wakatipu, with matai (*Prumnopitys taxifolia*) and kahikatea (*Dacrycarpus dacrydioides*) confined to the warmest localities. They also show that deforestation was extensive from *c*. 600 to 800 years ago (McGlone and Wilmshurst, 1999). Today, remnants of mountain beech (*Nothofagus solandri*) and silver beech (*N. menziesii*) forest grow in fire-sheltered pockets, and areas of mountain beech survive in wetter watersheds to the south. At the time of European arrival, snow tussock (*Chionochloa* spp.) grassland dominated at higher elevations and short tussock (*Poa cita, Festuca novae-zelandiae* and *Elymus apricus*) grassland prevailed in drier inland sites (McGlone, 2001). Most of these grasslands were converted to pasture by Europeans, and pine plantations were commercially widespread after the 1960s (Roche, 1990).

The northernmost study area (Lake Wanaka area) is generally warmer and drier than the other study areas (Table 1). Two lakes were cored: Diamond Lake and Glendhu Lagoon. Both landscapes currently support a mixture of agricultural land, primarily pasture, and small remnants of native vegetation. Open vegetation is dominated by bracken (*Pteridium esculentum*), matagouri (*Discaria toumatou*) and grasses (Poaceae). Forest remnants consist of small trees including *Coprosma* spp, *Olearia* spp, *Aristotelia serrata*, *Fuchsia excorticata*, *Griselinia littoralis*, *Pittosporum* spp. and *Pseudopanax* spp.

The Lake Wakatipu area is slightly cooler and wetter than the more northern Wanaka region (Table 1). It was part of a trade route for greenstone, a prized resource which Māori used and transported extensively around New Zealand (Chapman, 1891; Beattie, 1920; Brailsford, 1996; Central Index of New Zealand Archaeological Sites (CINZAS), 2008). Lake Kirkpatrick, the highest elevation site of the study, is surrounded by pasturelands with brown-

Table 1 Site locations, nearby meteorological site (Met. site), and 30 year mean climate observations

Site	Elev. (m)	Latitude (°S)	Longitude (°E)	Met. site	Elev. (m)	Latitude (°S)	Longitude (°E)	MAT (°C)	MinAT (°C)	PPT (mm/yr)
Lake Diamond	380	44.65	168.96	Wanaka	314	44.71	169.15	10.5	4.8	704
Glendhu Lagoon	280	44.66	169.05	Wanaka	314	44.71	169.15	10.5	4.8	704
Lake Kirkpatrick	570	45.03	168.57	Queenstown	329	45.04	168.66	10.5	5.5	851
Lake Te Aroha	290	45.33	167.80	Te Anau	204	45.43	167.72	9.7	4.7	1174
Lake Thomas	490	45.47	167.95	The Haycocks	385	45.50	169.00	9.7[a]	4.7[a]	1127

Mean annual temperature (MAT), mean annual minimum temperature (MinAT) and total annual precipitation (PPT) (1961–1990) were extracted from the New Zealand National Climate Centre database (1980–2008).

[a] MAT and MinAT data from the Haycocks were not available. The next nearest site with these data was the Te Anau met site.

top (*Agrostis capillaris*), *Poa* spp. and other introduced grasses, clover (*Trifolium* spp.), *Hypericum* spp., pine plantations (*Pinus radiata*) and small areas of native beech forest (*Nothofagus menziesii*).

The Lake Te Anau area is the farthest south and most inland study area and experiences colder and wetter conditions than the northern catchments (Table 1). Two lakes were cored: Lakes Thomas and Te Aroha. Lake Thomas is located in an agricultural setting dominated by grass and fern-shrubland, scattered tree plantations (*Cupressus, Eucalyptus, Pinus* spp.), small remnants of native shrubs and trees (*Pseudopanax colensi, Coprosma* and *Coriaria spp.*) and a small wetland on its northern margin. Native grassland is composed primarily of bracken (*Pteridium* spp.), tussock grasses (*Chionochloa* spp.), *Coprosma* spp. and occasional *Discaria* spp. Intact native beech (*N. solandri*) forests lie less than 5 km to the north and east of the watershed. Lake Te Aroha is primarily surrounded by extensive bogs dominated by *Sphagnum* and wirerush (*Empodisma minus*) with stunted scrub including manuka (*Leptospermum scoparium*), bog pine (*Halocarpus bidwillii*), *Dracophyllum* spp. and bracken (Wilmshurst *et al.*, 2002). A few patches of native forest surround the wetlands, dominated by beech (*N. cliffortioides*) but also including *Metrosideros umbellata, Weinmannia racemosa, Dracophyllum* spp. and *Phyllocladus alpinus* (Wilmshurst *et al.*, 2002).

Methods

Core collection

For the lake sites, short cores (between 0.79 and 1.69 m long) were recovered in 2007 from a floating anchored platform with a 1.75 m long, 7.5 cm diameter polycarbonate surface corer. This coring device preserves the mud–water interface intact. At Glendhu Lagoon, a 50 cm diameter peat core was recovered with a D-section corer in 1996. All lake cores were extruded and subsampled contiguously every 1 cm in the field and placed in plastic bags for the laboratory analyses described below. The peat core was extruded, wrapped in cellophane and transported intact back to the laboratory, refrigerated and subsequently sampled at 2 cm intervals.

Chronology

Plant macrofossils found in the cores were identified and placed in glass vials for radiocarbon dating. Age determinations were based on accelerator mass spectrometry (AMS) ^{14}C dates obtained on twig charcoal and plant macrofossils (twigs, leaves and plant fragments) for each of the five sites. We chose charcoal fragments associated with young or annual plant growth (eg, small twigs, leaves) in order to avoid dating older wood and any associated problems of in-built age (McFadgen, 1982; Gavin, 2001). Ages of all dated macrofossils are expressed as calibrated ^{14}C age before present (cal. yr BP), using Oxcal 4.0 (Bronk Ramsey, 1995,

2001) and the Southern Hemisphere (SHCal04) calibration curve (McCormac *et al.*, 2004). Timescales in the text and figures are expressed as cal. yr AD/BC.

Age–depth models were constructed based on the suite of calibrated ^{14}C macrofossil dates and an assigned age of AD 1970 at the depth where the most significant increase in exotic *Pinus* pollen (primarily *Pinus radiata*) was noted. Pine species were introduced to the South Island beginning in the late 1800s and were widespread but not abundant until large commercial plantations were established from the 1920s onwards (Roche, 1990). We utilize the most pronounced rise in pine pollen in our records as an indicator marking a second wave of plantations occurring in the latter 1900s. All ^{14}C macrofossil dates bracketing the initial increase in charcoal accumulation rates (CHAR, particles/cm^2 per yr) had errors that overlapped statistically when converted to cal. yr BP (Table 2). To identify the maximum duration of the fire episode, we constructed age–depth models constraining all calibrated ^{14}C macrofossil dates to occur sequentially in the sediment record (Figure 2). These age–depth models were derived using a multi-parameter Bayesian analysis Markov Chain Monte Carlo (MCMC) simulation model in Oxcal 4.0 (Bronk Ramsey, 1995, 2001, 2008). Thus, we used calibrated ^{14}C macrofossil dates (short duration or instantaneous) and dates from this constrained MCMC age–depth model (maximum duration) to determine the span of the initial burning period (IBP).

Lithological analyses

Magnetic susceptibility was analysed at all sites to measure increases in inorganic allochthonous sediment that could be attributed to erosion events (Gedye *et al.*, 2000). Subsamples of 5 cm^3 were taken for each 1 cm interval for the cores, placed into a plastic vial, and measured with a Bartington magnetic susceptibility meter cup sampler. Units are presented as centimetre-gram-second (CGS × 10^{-6}).

Organic and carbonate content was analysed to determine changes in lake production through time. Weight-loss after igniting dried samples at 550 and 900°C for 2 h determined the percent organics and percent carbonates, respectively (Dean, 1974). Measurements were made at 2 cm intervals for all lake sediment cores. Organic and carbonate content were not measured on the Glendhu Lagoon core.

Pollen analyses

Pollen data provide information on the regional and local vegetation history. At Lake Diamond and Lake Thomas, pollen was sampled every 50–100 years for the last 1000 years (17 samples at Diamond Lake and 39 samples at Lake Thomas). Samples were processed using methods of Bennett and Willis (2002). A *Lycopodium* tracer was added to calculate pollen concentration (grains/cm^3) and pollen accumulation rates (PAR) (grains/cm^2 per yr) were calculated

Table 2   Radiocarbon and pollen determined age information for study sites

| Site	Depth (cm)	Uncalibrated ^{14}C age (^{14}C yr BP)	Calibrated age (cal. yr BP)	Calibrated ^{14}C age (cal. yr BP)	Age (AD)	Material dated	Lab ID KCCAMS *CAMS
Diamond Lake	20	–	–	–	*c.* 1970	inferred from *Pinus* pollen	–
Diamond Lake	64	640±40	602±33	519±11	1431±11	plant fragments	39052
Diamond Lake	89	585±25	543±24	622±7	1328±7	wood charcoal	39053
Diamond Lake	101	820±20	702±16	704±15	1246±15	wood charcoal	42006
Glendhu Lagoon	80	540±25	524±11	548±18	1402±18	wood charcoal	42012
Glendhu Lagoon	74	–	–	–	*c.* 1430	inferred from Diamond Lake CHAR peak	–
Lake Kirkpatrick	8	–	–	–	*c.* 1970	inferred from *Pinus* pollen	–
Lake Kirkpatrick	95	690±20	599±31	569±8	1381±8	wood charcoal	42007
Lake Kirkpatrick	117	705±15	595±33	651±8	1299±8	wood charcoal	42008
Lake Te Aroha	33	535±20	522±10	502±8	1448±8	leaf	42011
Lake Te Aroha	47	510±15	512±7	534±6	1416±6	wood	39055
Lake Thomas	22	–	–	–	*c.* 1970	inferred from *Pinus* pollen	–
Lake Thomas	95	590±140	589±118	320±21	1630±21	wood charcoal	42009
Lake Thomas	138	410±30	482±53	503±14	1447±14	wood charcoal	42010
Lake Thomas	154	2500±45	2579±93	2361±86	BC 411±86	wood	*133155
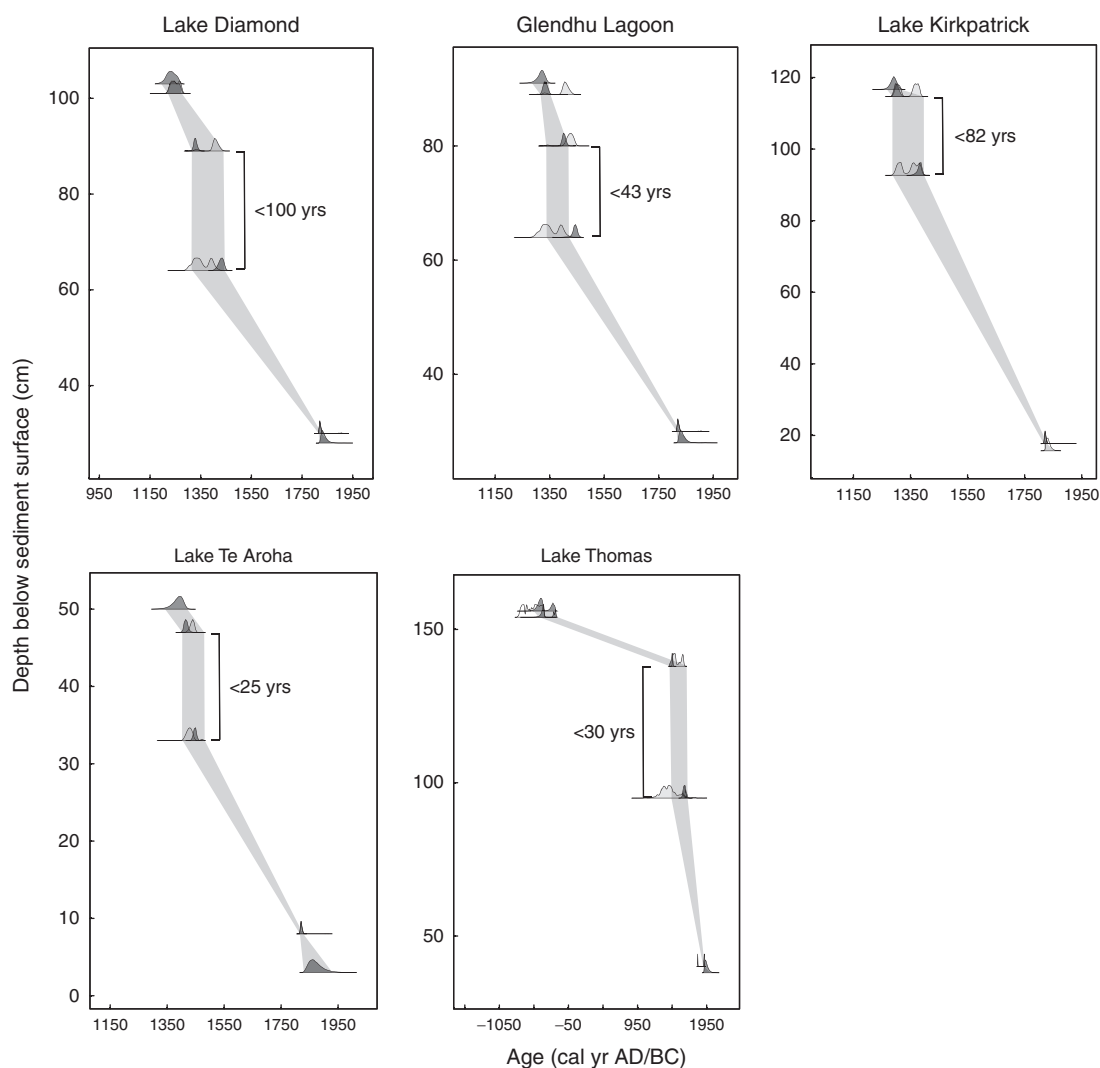
Figure 2   Age–depth models for each site constructed using a multiparameter Bayesian analysis Markov Chain Monte Carlo (MCMC) simulation model in Oxcal 4.0 (Bronk Ramsey, 1995, 2001, 2008). Light grey distributions indicate probability distributions from calibrated ^{14}C dates. Dark grey distributions indicate modelled probability distributions of possible ages based on Bayesian analysis MCMC. Bracketed period indicates overlapping ^{14}C dates. Maximum duration of bracketed interval (indicated by < xx yrs) derived by constraining ^{14}C dates to occur sequentially in the MCMC model

by dividing the sample by its deposition time (yr/cm). Pollen grains were identified at magnifications of 400 and 1000×, and sample counts ranged from 282 to 559 terrestrial grains/sample. Pollen was identified to the lowest taxonomic level possible with reference collections (Landcare Research), atlases (eg, Moar, 1993), and other publications (Pocknall, 1981a, 1981b). Pollen grains that were broken, corroded, hidden or otherwise damaged were counted as 'Indeterminate', and those that were unidentifiable were 'Unknown'. Pollen percentages were based on a sum including all trees, shrubs, herbs and bracken spores. Pollen percentages and PAR were used to reconstruct past vegetation.

Following pollen analyses, data were grouped into three categories: native forest taxa (primarily represented by families Podocarpaceae and Nothofagaceae); disturbance related taxa associated with post-fire succession (*Coprosma* spp., *Coriaria* spp., *Pteridium esculentum*, *Typha* spp. and family: Poaceae) and taxa introduced during European colonization of New Zealand: *Pinus* spp., *Trifolium* spp., *Rumex acetosella*. These groupings are used to identify the transition from closed-forest taxa associated with infrequent disturbance to post-disturbance taxa associated with anthropogenic burning and to identify key dates following European colonization associated with the presence of widespread *Pinus* plantations and the increase in pollen from introduced herbaceous species associated with pasture (eg, *Trifolium* spp., *Rumex acetosella*).

Charcoal analyses

High-resolution charcoal analysis following methods of Whitlock and Larsen (2001) was undertaken on all of the cores to reconstruct local fire activity. Sediment samples of 5 cm^3 were taken at contiguous 1 cm intervals and soaked in 5% sodium metaphosphate and 6% bleach for 24 h. They were washed through 125 µm mesh screens, and residues were counted in gridded Petri dishes. Charcoal particles were counted under a stereomicroscope at 50–100× magnification. Charcoal concentration (number of particles/cm^3) was calculated by dividing charcoal counts by the volume of sediment sieved. Charcoal accumulation rates (CHAR, particles/cm^2 per yr) were determined by dividing charcoal concentration by the deposition time (yr/cm) of each sample.

Concentrations and sedimentation rates were interpolated to pseudo-annual values and then binned in 7 yr long intervals approximating the median temporal resolution of each record. We describe the annual values as pseudo-annual values because the 7 yr intervals represent the finest resolution that is realistic for our records. Sedimentation rate may vary and bioturbation is likely, thus, it is necessary to choose an interval that approximates the median temporal resolution of the cores. This approach, which is now widely used in the palaeofire community (eg, Briles *et al*., 2008; Higuera *et al*, 2008; Marlon *et al*., 2008), maintains the original structure of the charcoal data, while avoiding pseudo-replication and smoothing of the data.

The CHAR time series was decomposed into two components: (1) a background component representing the long-term changes in deposition, and (2) a peak component representing the positive residuals after the background component was removed from the interpolated record. Decomposition of CHAR and statistical treatment of the charcoal records used the program CharAnalysis (Higuera, 2008; Higuera *et al*., 2008). The slowly varying component, often referred to as background or BCHAR, provided information on general characteristics of fires and fuel conditions (ie, area burned, regional fires, levels of available fuel biomass) (Marlon *et al*., 2006). Because we were interested in the details of the charcoal stratigraphy, we examined total CHAR data as opposed to the smoothed BCHAR output. Significant charcoal peaks, representing fire episodes, were identified as CHAR values greater than a locally defined threshold value. We used a Gaussian

mixture model to identify the mean and variance of the BCHAR distribution; the 99th percentile of this distribution was the threshold value separating peak fire episodes from 'noise'. A fire episode refers to one or more fires occurring within the time span of the charcoal peak. The statistical significance of each peak was evaluated by comparing the original charcoal counts with the values in samples occurring 35 years before the peak. If the maximum count of a peak had a >5% chance of coming from the same Poisson-distributed population as the minimum charcoal count within the preceding 35 years, then a 'peak' was not identified (Higuera, 2008).

At Diamond Lake and Lake Thomas, microscopic charcoal (charcoal particles < 100 µm) on pollen slides was also analysed as part of routine pollen counting to infer regional fire activity (see below). Microscopic particle counts were converted to charcoal concentrations (particles/cm^3) similar to the pollen data, then charcoal influx rates (particles/cm^2 per yr) were obtained by dividing charcoal concentration by deposition time (yr/cm). Hereafter we refer to microscopic charcoal accumulation rates as microscopic charcoal influx (particles/cm^2 per yr).

Reconstruction of the fire history

Fire-history reconstructions based on lake-sediment records were derived from (1) macroscopic charcoal accumulation rates that provide direct evidence of local fire; (2) microscopic charcoal influx rates that describe regional or general patterns in fire activity; (3) pollen evidence of changes in plant communities related to disturbance; and (4) lithologic and geochemical evidence of watershed responses to fire, such as increased erosion.

The interpretation of high-resolution charcoal records as a local fire proxy is based on empirical studies following modern fires; modelling studies that consider charcoal production, transport and deposition; and statistically rigorous efforts to discriminate between signal and noise in the charcoal time series. The distance that charcoal is carried during a fire has been discussed in several papers (see Clark and Royall, 1995; Whitlock and Larsen, 2001; Gavin *et al*., 2003; Higuera *et al*., 2007). Simple Gaussian plume models suggest that particles >100 µm diameter are released relatively close to the ground and deposited in close proximity to a fire, whereas particles < 100 µm diameter in size travel well beyond 100 m, and very small particles are capable of being lofted to great heights and travel long distances (Clark and Patterson, 1997). Thus, macroscopic charcoal provides a signal of fire activity near the depositional site.

In small lakes with minimal stream inflow and within small watersheds, such as those analysed in this study, variations in macroscopic charcoal accumulation rates (CHAR) describe local fire patterns. Trends in BCHAR are related to changes in vegetation and fuel biomass (Marlon *et al*., 2006), whereas peaks in CHAR (particles/cm^2 per yr) represent fires occurring within a few kilometres radius of the lake (Gavin *et al*., 2003; Higuera *et al*., 2007). Peak magnitude (particles/cm^2 per peak), the charcoal flux of a given peak, is used as a measure of fire size or severity (ie, charcoal production) (Whitlock *et al*., 2006; Higuera *et al*., 2007, 2008). Nearly all of the particles in our study are burned wood and leaf fragments, suggesting that they came from high-severity fires. Such fires produce large charcoal particles and the convectional activity associated with them carries large particles aloft. In contrast, these data do a relatively poor job of reconstructing low-intensity or surface fires (Whitlock *et al*., 2004), which produce few or small particles.

Generally, the first significant increase in macroscopic CHAR (ie, the beginning of the charcoal peak) is used as the fire-episode date, and the peak represents one or more fires occurring during its total time span. Local fires introduce macroscopic charcoal to a lake via airborne fallout, but the input of significant levels of secondary charcoal can occur a few years after the fire, before slopes are stabilized by vegetation and as particles in the littoral

zone are redeposited to deep water (Whitlock and Millspaugh, 1996). Subsequent bioturbation of the sediments by benthic organisms may also lengthen the time span of the charcoal peak, extending the signal by years in lakes with slow accumulation rates.

Microscopic charcoal can travel long distances and in New Zealand some of this particle size may come from fires from Australia or Southeast Asia (Butler, 2008), or ground fires that produce little charcoal. Microscopic charcoal is evident in sediment cores from New Zealand throughout the Holocene, but the lack of an associated change in pollen composition suggests that these fires were distant (Molloy *et al.*, 1963; McGlone *et al.*, 1992, 2004) or caused relatively short-term and minor vegetation changes (Wilmshurst and McGlone, 1996; Wilmshurst *et al.*, 1997). Our attention is the large increase in fire activity that occurred soon after first known Māori presence. We term this period the Initial Burning Period or IBP.

Evidence of watershed erosion and limnobiotic change

Measurements of magnetic susceptibility, organic and carbonate content are commonly used to record changes in watershed conditions over time. Peaks in magnetic susceptibility, an indicator of deposition of inorganic material into the lake, often signal increases in erosion or an erosion event where a significant amount of mineral soils were deposited rapidly (Gedye *et al.*, 2000). These events are often associated with disturbances such as fire. Changes in organic content in lake sediment can be used as a surrogate for the productivity of the lake or watershed, whereas abrupt decreases often signal the dilution of organic material with the input of inorganic material through erosion events (Dean, 1974). Carbonate content in lake sediments serves as an indicator of changes in either the input of carbonate-rich sediment from terrestrial sources via erosion or changes in lake geochemistry that influence detrital contributions of carbonates (Dean, 1974). Hence, magnetic susceptibility and organic and carbonate content can be used to infer erosion events, changes in lake productivity and changes to the lake environment associated with fire episodes and vegetation change evident in charcoal and pollen records.

Results

Lake Wanaka area
Lithology and chronology
The Diamond Lake core was characterized by fine-detritus gyttja (nutrient-rich organic mud) with little stratigraphic variation. The Glendhu Lagoon core consisted of partially decomposed peat from wetland vegetation dominated by *Typha*, and the only obvious change in core lithology was a layer of mixed silt and sand at 2–3 cm depth. Pronounced changes in sedimentation were identified at 88–89 cm (0.13 to 0.38 cm/yr), 64–65 cm (0.38 to 0.08 cm/yr) and 24–25 cm depth at Diamond Lake (0.08 to 0.37 cm/yr) and at 80–81 cm depth at Glendhu Lagoon (0.11 to 0.37 cm/yr).

Loss-on-ignition and magnetic susceptibility
The organic content of the Diamond Lake core gradually increased from 40% at *c.* AD 1000 to > 50% at *c.* 1900s, with one significant, yet brief, decrease in organic content from 40 to 18% associated with the IBP at *c.* AD 1420 (Figure 3). Percent carbonate content was low (2–5%), with an increase from 5 to 10% in the late 1900s. Following the IBP (*c.* AD 1420), magnetic susceptibility tripled from 4 to >12 CGS × 10^{-6}. Magnetic susceptibility then decreased in the late AD 1400s to 3–4 CGS × 10^{-6} and remained at these levels until present.

At Glendhu Lagoon, magnetic susceptibility was negative or near zero until the top of the core, at *c.* AD 1940, where it increased from 0 to 5 CGS × 10^{-6}, then decreased to 0 again at *c.* AD 1980. Loss-on-ignition was not analysed for Glendhu Lagoon.

Charcoal
At Diamond Lake, CHAR was extremely low (< 1 particles/cm^2 per yr) until *c.* AD 1340 when CHAR increased to 19.09 particles/cm^2 per yr. This initial peak was then followed by a second peak of 22.89 particles/cm^2 per yr at *c.* AD 1420. CHAR decreased dramatically (< 1 particles/cm^2 per yr) until *c.* AD 1850 when it rose slightly to 1.03 particles/cm^2 per yr and then again to 5.14 particles/cm^2 per yr at *c.* AD 1980. Significant fire episodes were identified at *c.* AD 1340, 1420, 1850 and 1980. Microscopic charcoal influx followed a similar pattern to macroscopic CHAR, increasing abruptly to > 35 000 particles/cm^2 per yr at *c.* AD 1328, decreasing to low levels (<1000 particles/cm^2 per yr) after the IBP and increasing again in the mid-nineteenth century.

Similar to the record at Diamond Lake, CHAR at Glendhu Lagoon was low (< 1 particles/cm^2 per yr) until *c.* AD 1390, after which it rose to 1.26 particles/cm^2 per yr. CHAR values increased at *c.* AD 1430 to 3.27 particles/cm^2 per yr and again at *c.* AD 1740 to 1.47 particles/cm^2 per yr and rose to maximum values, 15.98 particles/cm^2 per yr, at *c.* AD 1900. Fire episodes were identified at *c.* AD 1390, 1580, 1740, 1900 and 1960.

Pollen
Pollen data from Diamond Lake show a dramatic shift from native forest taxa: (*Podocarpus*, *Prumnopitys taxifolia* and *Nothofagus menziesii*) to shrubs (*Coprosma* spp., *Coriaria* spp.), bracken (*Pteridium esculentum*), grasses (family: *Poaceae*) and *Typha orientalis*, Figure 3). Pollen of *Podocarpus* (57%) and *Nothofagus* (12%) decreased to 5.6 and 6%, respectively, following the IBP between AD 1340 and 1420, whereas low percentages of *Coprosma* (6%), *Pteridium esculentum* (<1%), and Poaceae (< 1%) increased to 12, 55 and 16%, respectively. Percentages of native tree taxa remained low, and those of shrubs, bracken and grass were high to the present day. Following an initial peak of 55%, bracken decreased to 26–28% until AD 1700, after which it increased to as high as 35%. *Typha* pollen levels increased from <1% to > 40% at *c.* AD 1400. This increase is noteworthy because the presence of *Typha* implies increased inwash of silt, sand and nutrients associated with watershed disturbances, such as fire (McGlone and Wilmshurst, 1999). Pollen from *Pinus* spp., *Rumex acetosella* and other exotic species first appeared in the 1900s, and reach highest percentages in the late 1900s.

Lake Wakatipu area
Lithology and chronology
The Lake Kirkpatrick core was composed of fine-detritus gyttja with little lithologic variation but a change in colour at 118 cm depth. Radiocarbon dating and the chronology for Lake Kirkpatrick show that sedimentation rates increased during the initial period of increased CHAR from 0.14 to 0.27 cm/yr at 94 cm depth and again from 0.14 to 0.41 cm/yr at 14 cm depth.

Loss-on-ignition and magnetic susceptibility
Magnetic susceptibility levels remained between 3 and 4 CGS × 10^{-6} with two distinct increases: an abrupt rise to 8.8 CGS × 10^{-6}, coincident with the initial increase in CHAR at *c.* AD 1300 and a second more gradual increase to levels above 5 CGS × 10^{-6} (peaking at 7 CGS × 10^{-6} in AD 1804) for more than a century from *c.* AD 1780 to 1880 (Figure 4). This second rise occurred during a period with two large fire episodes at *c.* AD 1780 and 1850.
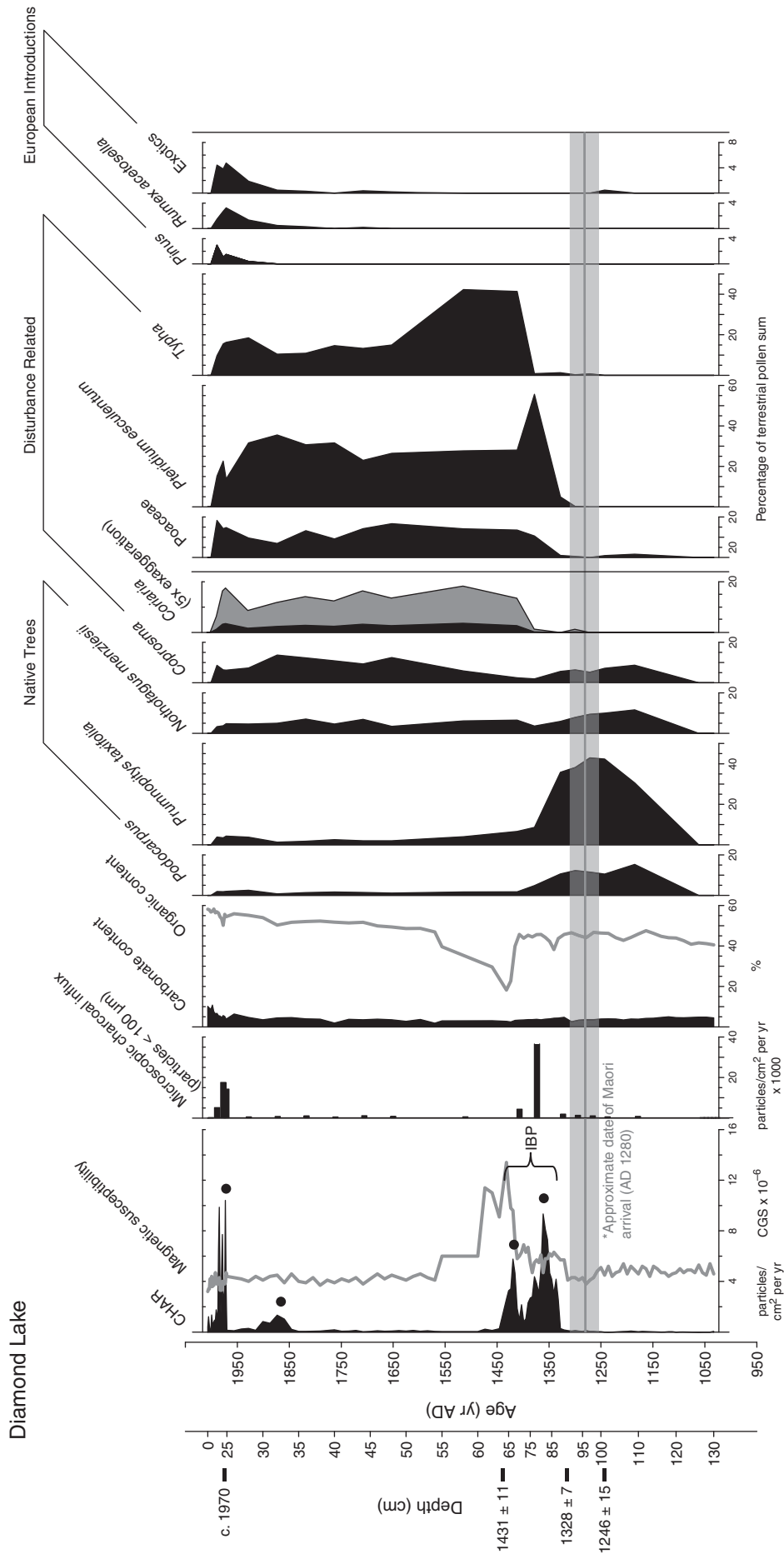
Figure 3 CHAR, fire episodes, magnetic susceptibility and selected pollen percentages for the Lake Diamond site. Black circles represent fire episodes identified from decomposition analysis. Grey line and shading indicate approximate date of Māori arrival (Wilmshurst *et al.*, 2008)
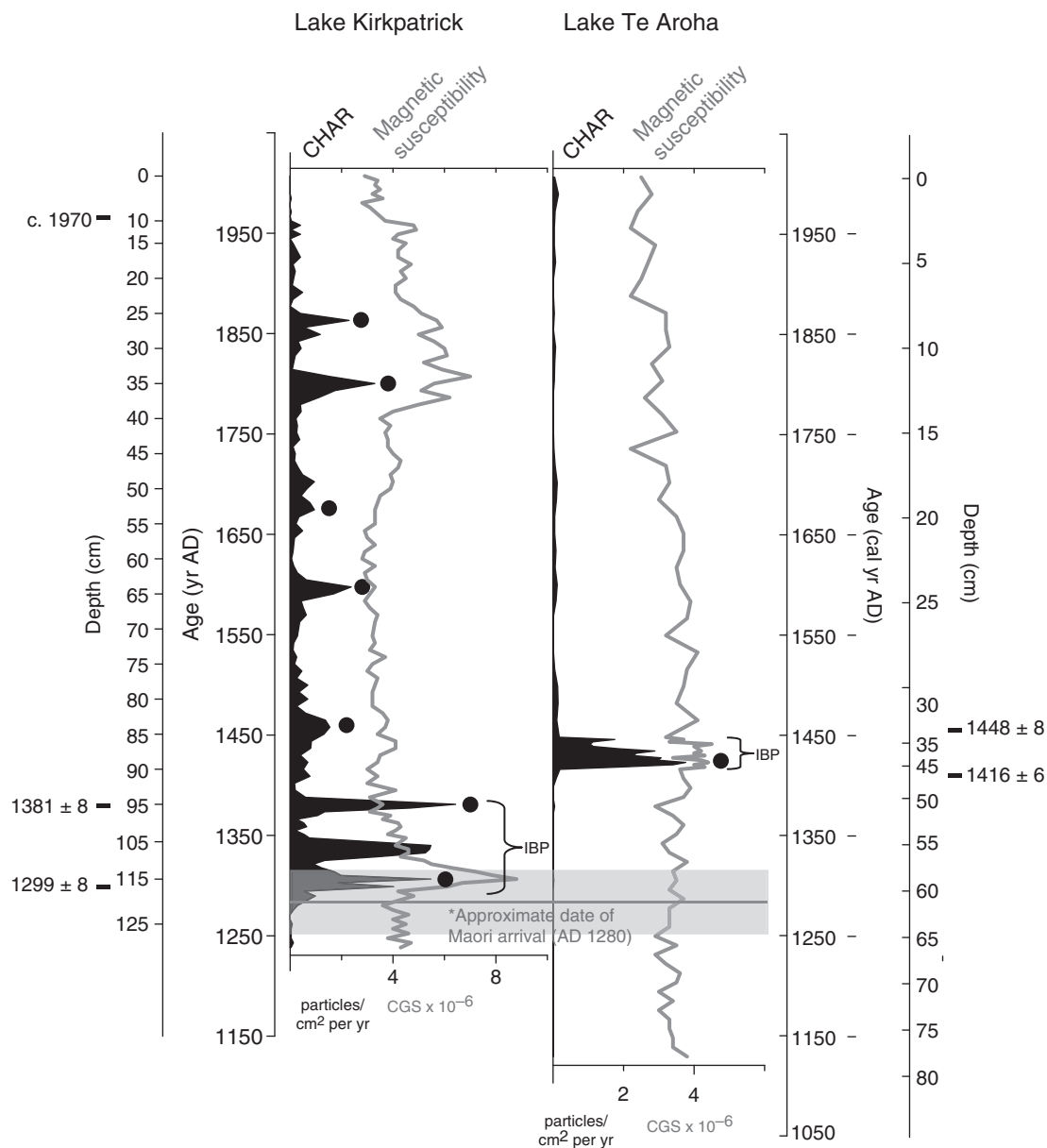
Figure 4 CHAR, fire episodes and magnetic susceptibility for the Lake Kirkpatrick and Lake Te Aroha sites. Black circles represent fire episodes identified through decomposition analysis

Charcoal record

The IBP was punctuated by two fire episodes with CHAR of 3.8 particles/cm² per yr at *c.* AD 1300 and 3.27 particles/cm² per yr at AD 1384. A sharp peak in magnetic susceptibility suggests that this interval may have occurred more quickly (within several decades) than the MCMC model chronology would suggest (<100 yr). Following the initial increase in CHAR, fire episodes occurred every 100–150 yr until the twentieth century when fire activity appeared to decrease. Five fire episodes were recorded at *c.* AD 1450, 1590, 1680, 1810 and 1860.

Lake Te Anau area
Lithology and chronology

Lake Thomas and Lake Te Aroha cores were composed of fine-detritus gyttja with little variation in lithology or colour. At Lake Thomas, a ¹⁴C date from a macrofossil located at 145–146 cm depth was calibrated at *c.* 560 BC, suggesting that sedimentation rates were slow (< 0.02 cm²/yr) prior to known Māori presence. Sedimentation rates increased at both sites in association with the initial increase in CHAR, from 0.01 to 0.24 cm/yr at 137 cm depth

at Lake Thomas and from 0.11 to 0.44 cm/yr at 46 cm depth at Lake Te Aroha. At Lake Thomas, a second period of increased fire activity from *c.* AD 1950 to present is also associated with high sedimentation rates.

Loss-on-ignition and magnetic susceptibility

Magnetic susceptibility levels at Lake Thomas remained between 2.5 and 3.5 CGS × 10⁻⁶ with two distinct increases associated with high CHAR values (Figure 5). Magnetic susceptibility increased from 3.3 to 5.3 CGS × 10⁻⁶, then remained between 4 and 5 CGS × 10⁻⁶ for almost 200 yr before dropping to levels between 2.5 and 3.5 CGS × 10⁻⁶. A second increase occurred in the late 1900s (to 4.1 CGS × 10⁻⁶) after the last fire episode at *c.* AD 1950. Organic content showed little variation throughout most of the record, ranging between 47 and 51%, but percentages rose slightly from 51 to 57% at the top of the core from *c.* 1990 to present. Carbonate content showed a pronounced increase from low levels (<7%) after the initial increase in CHAR at *c.* AD 1450, stayed above 10% for more than 100 yr (peaking at 35% at *c.* AD 1490), then decreased (2–7%) between *c.* AD 1550 and present.
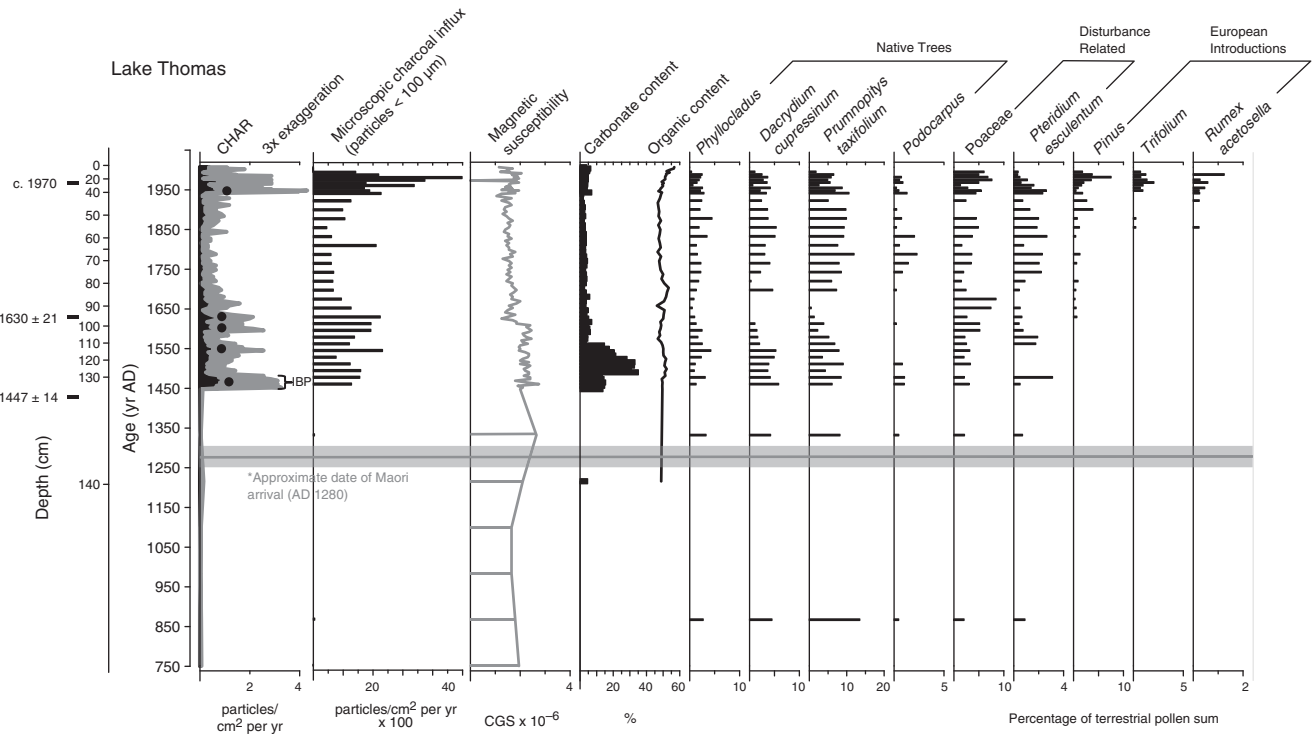
Figure 5   CHAR, fire episodes, magnetic susceptibility, charcoal concentration, carbonate content, organic content and selected pollen percentages for the Lake Thomas site. Black circles represent fire episodes identified through decomposition analysis

Only one peak in magnetic susceptibility was apparent at Lake Te Aroha, occurring at the end of the initial rise in CHAR and in association with a fire episode at AD 1419 (Figure 4). Magnetic susceptibility rose to levels > 4 CGS × 10^{-6} during this period and peaked at 4.5 CGS × 10^{-6} at AD 1440. Percent organic content varied between 60 and 80%, showing one noticeable decrease to 47% at AD 1800. Percent carbonate content showed little change at Lake Te Aroha, remaining between 4 and 10%.

Charcoal record
Charcoal data from Lake Thomas showed an initial increase in CHAR at AD 1450 from > 0.01 to 0.92 particles/cm² per yr (Figure 5). This initial rise continued until the early 1600s with fire episodes at *c.* AD 1550, 1600 and 1630. CHAR rose in the mid 1900s with a fire episode recorded AD 1951. Despite relatively low charcoal concentrations and little variation in CHAR, five significant fire episodes were identified in the record at *c.* AD 1450, 1550, 1600, 1630 and 1950. Values of microscopic charcoal influx changed similarly to those of macroscopic CHAR.

Lake Te Aroha, approximately 19 km northwest of Lake Thomas, recorded only two rises in CHAR associated with significant fire episodes, one initial rise in CHAR from levels < 0.01 to 3.69 particles/cm² per yr at *c.* AD 1440 (coinciding with the timing of an increase in charcoal concentrations at an adjacent bog, Wilmshurst *et al.*, 2002) and a second increase from similarly low levels to 0.15 particles/cm² per yr at *c.* AD 1980 (Figure 4).

Pollen record
The pollen record from Lake Thomas showed a gradual decrease in the percentages of two tree species (*Dacrydium cupressinum* and *Prumnopitys ferruginea*) and an increase in *Pteridium esculentum* and *Poaceae*, Figure 5). Pollen of *D. cupressinum* and *P. ferruginea* decreased from 7 and 10% to 0% during the interval from *c.* AD 1450 to 1670, while bracken and grass values increased from 2 and < 1% to 10 and 3%. Percentages for *Podocarpus* showed a similar trend.

This change coincided with a period of increased CHAR and microscopic charcoal influx and four local fire episodes. Percentages of native forest taxa recovered in the AD 1700s, decreased for a second time at *c.* AD 1900 then increased modestly in the mid-twentieth century. Exotic plant species (*Rumex* and *Trifolium* spp.) pollen were present (1.5 and 2%) in the late AD 1900s.

Discussion

Results from the five watersheds in this study provide new information on the timing, duration and frequency of fires associated with the Initial Burning Period (IBP) and the ecological response to the IBP including changes in vegetation, slope stability and limnology. Our five sites were compared with charcoal data from two bogs: Travis Swamp, located in the eastern central coast of the South Island on the outskirts of Christchurch, and Pomahaka, located approximately 90 km northwest of Dunedin, inland from the southeast coast of the South Island (Rose, 2004) to examine regional patterns in fire history (Figure 1 for site location). To investigate the relationship between climate and the timing of the IBP we utilize a 4000 yr long palaeoclimate reconstruction for New Zealand, based on multiple proxies (Lorrey *et al.*, 2008). This climate record allows us to compare the timing and patterns of fire at all sites with centuries of wet-versus-dry conditions in southwestern South Island.

The Initial Burning Period (*c.* AD **1280–1600**)
Timing of the IBP
Māori were known to be present in New Zealand at *c.* AD 1280 ± 30 (Wilmshurst *et al.*, 2008), as determined by radiocarbon ages of seeds gnawed by the introduced commensal Pacific rat (Figure 6). All of our sites show a dramatic increase in charcoal at around this time, although the signal is diachronous. It is possible that Travis Swamp and Pomahaka burned earlier (*c.* AD 1270–1280) than cooler, wetter, inland sites (*c.* AD 1400–1450 at Lake Te
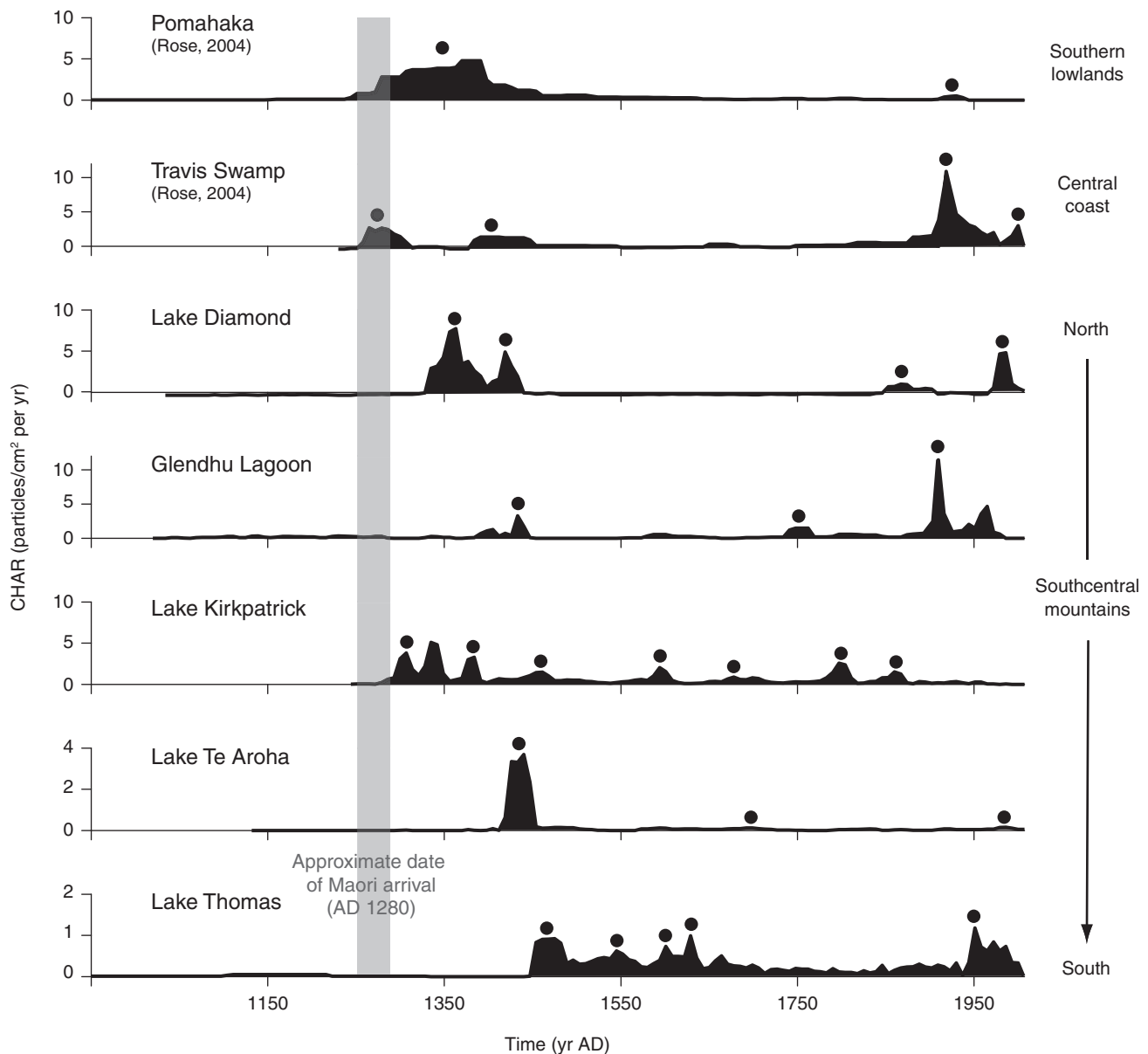
Figure 6 Comparison of charcoal records. CHAR and fire episodes for each site in this study and two additional sites from a previous study (Rose, 2004). Black circles represent fire episodes identified through decomposition analysis

Aroha and Lake Thomas). The IBP at Lake Kirkpatrick occurred soon after (*c*. AD 1290), and as much as 180 years later at sites further inland from the east coast. The temporal variation suggests that Māori fires may have cleared some areas earlier than others, but additional contiguous charcoal records and detailed radiocarbon dating are needed to elucidate the geographic pattern.

Duration and frequency of fires during the IBP
The duration of the IBP lasted from years to decades, depending on whether overlapping dates or the stratigraphically constrained Baysian age models are used to construct the chronology (Figure 7). The Baysian age–depth models suggest that the maximum duration of the IBP was less than a century at Diamond Lake (<100 yr) and Lake Kirkpatrick (<82 yr) and shorter at Glendhu Lagoon (<43 yr), Lake Thomas (<30 yr) and Lake Te Aroha (<25 yr). The plateau in the radiocarbon calibration curve associated with *c*. AD 1300–1400 makes it difficult to resolve whether the IBP was characterized by multiple fires or a single large conflagration. Multiple charcoal peaks were detected at Travis Swamp, Diamond Lake, Lake Kirkpatrick and Lake Thomas

(Figure 6), however, suggesting the possibility that repeated fires occurred over a short duration.

Ecological response to the IBP
Abrupt changes in vegetation (inferred from the pollen data), clastic sediment input (as indicated by sediment magnetism) and lake production (inferred from the carbon and carbonate content) were associated with the IPB. Pollen studies from numerous locations east of the Southern Alps show extensive deforestation throughout the South Island as a result of the IBP (McGlone, 1983, 1989; McGlone and Wilmshurst, 1999), occurring first at coastal and drier sites (McGlone and Wilmshurst, 1999; McGlone, 2001; McGlone *et al*., 2005). The predominant pattern indicates a transition from closed-canopy forests to bracken and early-successional shrubs (eg, *Coriaria*) and grasses which are well adapted to post-fire environments (McGlone *et al*., 2005).

As indicated from other studies, fire has a greater impact on closed forests when it is followed by a series of reburns (Uhl and Kauffmann, 1990; Cochrane et al., 1999; Cochrane, 2003; Thompson *et al*., 2007). It seems improbable that such forest loss would
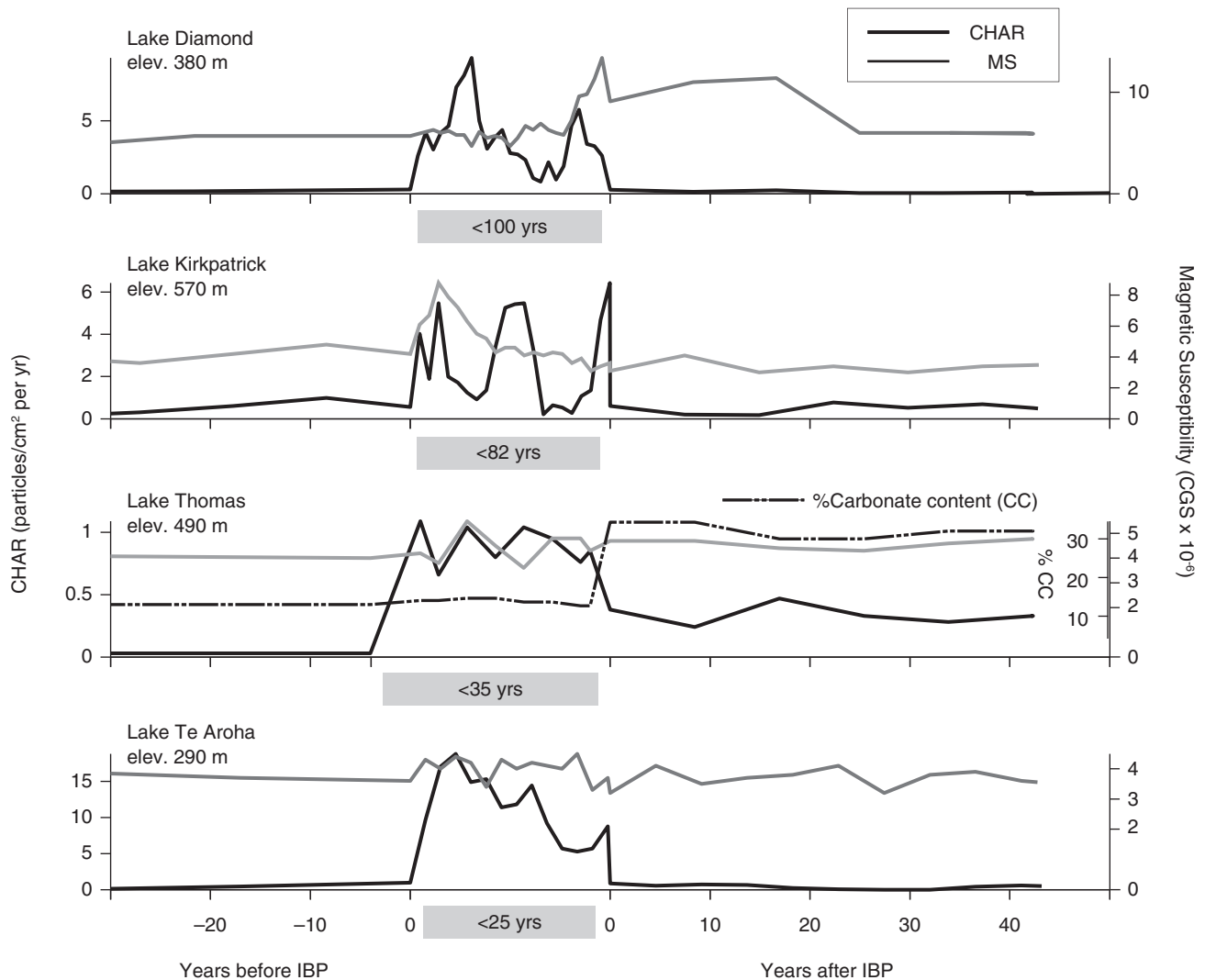
Figure 7   CHAR, magnetic susceptibility (MS) and percent carbonate content (Lake Thomas only) for four lakes, before, during and after the IBP. Maximum duration of the IBP indicated below each interval of increased CHAR. Chronologies for each site are based on a Baysian age–depth model where radiocarbon dates are constrained to occur sequentially

occur in a single event, and we suggest that the watershed conversion of native forests to grass and fern-shrublands was a consequence of several fire episodes over a short time interval.

Evidence of erosion during the IBP is evident at Lake Kirkpatrick, Lake Thomas and Lake Te Aroha, and within a decade of the IBP at Diamond Lake (Figure 7). The increased soil erosion following the IBP was relatively short-lived (see also Wilmshurst *et al*., 1999), and soil strength and cohesion probably improved with the establishment of bracken. Bracken maintains soil structure with its deeply penetrating and fire-proof rhizomes, and protects the soil surface from rain with its thick litter layers and dense canopy (Wilmshurst *et al*., 1997; McGlone *et al*., 2005). In addition, intact roots of burnt trees would have continued to provide significant soil cohesion until they decayed. In the closed forests of the northwestern North America, the potential for mass-wasting events has been shown to increase for several years after deforestation while dead roots decay and roots of new vegetation become fully established (Swanson, 1981). This may explain why erosion occurred later at Diamond Lake (Figure 7).

Changes in lake chemistry were most evident at Lake Thomas where carbonate content rose dramatically during the IBP and remained at elevated levels for several decades (Figure 5). This change towards more alkaline conditions suggests the first Māori

fires in the watershed triggered either increased input of carbonates from terrestrial sources via erosion or a change in limnobiotic conditions that influenced detrital input (eg, flux in algal populations) from within the lake environment. Further analysis of indicators of lake geochemistry may help resolve the source of this change at Lake Thomas. The increase in pollen from *Typha* spp. at Lake Thomas, an indicator of increased inwash of silt, sand and nutrients suggests further limnobiotic changes resulting from fires during the IBP (McGlone and Wilmshurst, 1999).

The influence of climate on the timing of the IBP
Natural fires in New Zealand require a successful ignition (from lightning or volcanic activity) and climatic conditions favourable for fire spread (warm temperatures and low humidity). This coincidence was apparently very infrequent during most of the Holocene, occurring every one to two millennia (Ogden *et al*., 1998). Lorrey *et al*. (2008) suggest the southwestern South Island was warm and dry from AD 1050 to 1200, mild and wet between AD 1250 and 1350, and mild and dry between AD 1400 and 1500. The warm and dry period precedes the IBP at Travis Swamp, Pomahaka, Lake Kirkpatrick and Diamond Lake, and the IBP occurred during the mild and wet period for these sites. In contrast, mild and wet conditions prevailed just prior to the IBP at Glendhu Lagoon, Lake Te Aroha and Lake Thomas

while the IBP occurred during a mild and dry period for these sites (Cook *et al*., 2006; Lorrey *et al*., 2008).

Although all of these data sets have chronological uncertainty, the inconsistent relationship between periods of aridity and the range of dates marking the IBP suggests that long-term climatic trends were not the primary driver of the IBP. A tree-ring reconstruction of summer temperature for the central west coast of the South Island shows considerable interannual variability during the past millennium (Cook *et al*., 2006). When burning, Māori would probably have taken advantage of dry seasons or years, even during periods when the long-term trend was towards wetter conditions. Large severe fires may also have been aided by fuel manipulations, including preparation of understorey and small trees as starter fuels (McGlone, 1983), as occurs in mesic forests elsewhere (Uhl and Kauffmann, 1990; Thompson *et al*., 2007).

The Late Māori period (*c.* AD 1600–1850)

Low levels of macroscopic and microscopic charcoal in the period between the IBP and European settlement indicate the presence of small or distant fires during this time. The reduction in CHAR after the IBP is attributed to decreased levels of woody fuel following the IBP, associated with an increase of seral taxa and little recovery of native forest species, especially at Diamond Lake (Figures 3, 5). Forests that had developed in the face of long fire-return intervals may have been slow to recover following the IBP (McQueen, 1951; Wardle, 1980, 1984; Wiser *et al*., 1997; Ogden *et al*., 1998). In contrast to rapid rates of forest regeneration following recent fires in *Nothofagus* forests (eg, Wiser *et al*., 1997), the slow rates of recovery recorded in the pollen diagrams during the Māori period suggest continued use of fire to maintain seral vegetation. Thus, the low levels or absence of CHAR following the IBP more likely reflects the fact that herbaceous and bracken-dominated vegetation do not produce much macroscopic charcoal when they are burned frequently.

At Lake Kirkpatrick, fire activity was sustained at intervals of 50 to 100 yr until European arrival in the mid-nineteenth century. Māori were known to have obtained greenstone at sites in the Lake Wakatipu area (Chapman, 1891; Beattie, 1920; Brailsford, 1996), and abundant archaeological evidence here suggests that the area was frequently travelled (Brailsford, 1996; CINZAS, 2008). Burning would have facilitated ease of travel and also promoted bracken as a food resource (McGlone *et al*., 2005). This use of fire may explain the numerous (7) fire episodes recorded at Lake Kirkpatrick and not at others. Alternatively, if frequent, low-severity burning of bracken at the other sites explains the lack of a charcoal signal there, the record at Lake Kirkpatrick may indicate less frequent but deliberate burning to remove encroaching woody species, such as manuka (*Leptospermum scoparium*) or kanuka (*Kunzea ericoides*).

The European period (*c.* AD 1850–present)

European arrival in the mid-nineteenth century led to a second period of burning and additional changes in vegetation that are recorded in the charcoal and pollen records. Macroscopic CHAR increases after AD 1850, and large-magnitude fire episodes at Diamond Lake, Glendhu Lagoon and Lake Thomas suggest increased fire activity into the twentieth century (Figure 6). Fire activity in the nineteenth century was less pronounced at Lake Kirkpatrick and Lake Te Aroha, perhaps because these areas were already largely deforested or, in the case of Lake Te Aroha, less useful for pasture because of poor soils. Smaller magnitude fires at these two sites could also have resulted from burning of herbaceous vegetation that produces little charcoal.

Records of early European settlement refer to the use of widespread fires to facilitate conversion of forest, fern-shrubland and tussock grasses to pasture forage for livestock (Clark, 1949;

Murphy, 1951; O'Connor *et al*., 1987; Guthrie-Smith, 1999). A number of introduced pasture plants, including *Trifolium* spp., *Rumex acetosella* and members of the Poaceae family, appear in the pollen record at Diamond Lake and Lake Thomas, especially in the late twentieth century. *Pinus* species were also introduced to New Zealand following European colonization, and plantations of *P. radiata* were established throughout New Zealand in the mid to late twentieth century (Roche, 1990). The magnitude of charcoal peaks during the European period is smaller than during the IBP at Pomahaka, Diamond Lake, Lake Kirkpatrick and Lake Te Aroha, and this likely reflects the lower amounts of woody material available from previous conversion of closed-forests to successional vegetation assemblages dominated by shrubs and grasses. European fires produced larger magnitude CHAR peaks at other sites (Travis Swamp, Glendhu Lagoon and Lake Thomas) than during the IBP. The motivations for European fires were to burn remnant forests, create young growth in the tussock cover and eliminate shrub patches that occupied potential grazing land. The reason charcoal accumulation rates increased during the European period at these sites is most likely because Europeans burned forest remnants at the perimeter of early-seral vegetation targeting shrubby species (with ample woody material) to increase pastureland. Thus, the type of vegetation that was targeted for burning (herbaceous versus woody) likely accounts for the variation in CHAR across the seven sites during the European period.

Conclusions

Our study contributes to the ongoing discussion of prehistoric and recent human impacts on New Zealand's vegetation (Sutton, 1994; McGlone and Wilmshurst, 1999; McGlone *et al*., 2005; Sutton *et al*., 2008). We provide the first application of high-resolution macroscopic charcoal analysis to document the fire history of New Zealand, and compare the local fire history with other proxies of environmental change. The study sites were located in small catchments with minimal inflowing streams, which enabled reconstruction of local fire activity. Because macroscopic charcoal peaks (ie, fire episodes) come from fires within a few kilometres radius of the lakes, a number of important conclusions follow from our results. First, to the best of our ability to date the event, the IBP was diachronous, occurring within 180 years among our sites between *c*. AD 1270 and 1450. We confirm what others have reported (McGlone and Wilmshurst, 1999; McGlone, 2001), namely that burning at inland sites occurred later than on the coast, and wetter remote sites were the last to be burned, but the geographic pattern of burning is still poorly known. Second, while the determination of the time span of the IBP is limited by the precision of radiocarbon dating, our chronologies suggest that the event lasted years to decades at each site and may have consisted of multiple reburns. This period was the major deforestation event in the history of each watershed, and it was accompanied by a dramatic transformation in vegetation, slope stability and limnology. These watershed responses occurred during or immediately after the IBP. Third, many sites show no fires after the IBP, but it is likely that our charcoal records do a poor job of registering small, frequent fires in vegetation dominated by herbaceous species such as bracken which produce little macroscopic charcoal. Subsequent fires, if present, were limited in scope yet may have prevented the recovery of the fire-sensitive woody species. Fourth, the one site that shows evidence of fire episodes after the IBP was located near greenstone sources, and periodic burning may have helped maintain travel routes to those areas. Alternatively, differences in post-IBP fire history may be associated with local site differences. For example, some sites may have had a higher percent of highly flammable species amongst the bracken fern-shrubland, such as

Phyllocladus alpinus, *Podocarpus* spp. and *Dracophyllum*, whereas others may have lacked sufficient fuels to support post-IBP fires. Fifth, the IBP cannot be easily attributed to long-term variations in climate, in particular moisture, occurring on centennial timescales. In fact, comparison with regional networks of palaeoenvironmental data suggests that some of the sites burned during wet centuries and others during dry centuries. More detailed analyses are needed to more thoroughly examine the relationship between climate and the timing and structure of the IBP. The weak relationship between long-term climate change does not rule out, however, the possibility that fires occurred during a dry season or year embedded in these century intervals, or that deliberate fuel manipulation may have enhanced fire intensity.

Additional high-resolution fire reconstructions are needed across a variety of watersheds, elevations and forest types to reconstruct the spatial and temporal pattern of fire activity. Improvements in the chronologies through a 'wiggle-match' ^{14}C dating approach that considers dates from contiguous samples in the same record will also help constrain the duration of the IBP (Kilian *et al.*, 2000; Speranza *et al.*, 2000; Hogg *et al.*, 2003). Finally, independent palaeoenvironmental proxy will help us evaluate the response of watershed changes in different settings to the IBP as well as past climatic variations.

Acknowledgements

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