

# Rapid landscape transformation in South Island, New Zealand, following initial Polynesian settlement

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Humans have altered natural patterns of fire for millennia, but the impact of human-set fires is thought to have been slight in wet closed-canopy forests. In the South Island of New Zealand, Polynesians (Māori), who arrived 700–800 calibrated years (cal y) ago, and then Europeans, who settled ~150 cal y ago, used fire as a tool for forest clearance, but the structure and environmental consequences of these fires are poorly understood. High-resolution charcoal and pollen records from 16 lakes were analyzed to reconstruct the fire and vegetation history of the last 1,000 y. Diatom, chironomid, and element concentration data were examined to identify disturbance-related limnobiologic and biogeochemical changes within burned watersheds. At most sites, several high-severity fire events occurred within the first two centuries of Māori arrival and were often accompanied by a transformation in vegetation, slope stability, and lake chemistry. Proxies of past climate suggest that human activity alone, rather than unusually dry or warm conditions, was responsible for this increased fire activity. The transformation of scrub to grassland by Europeans in the mid-19th century triggered further, sometimes severe, watershed change, through additional fires, erosion, and the introduction of nonnative plant species. Alteration of natural disturbance regimes had lasting impacts, primarily because native forests had little or no previous history of fire and little resilience to the severity of burning. Anthropogenic burning in New Zealand highlights the vulnerability of closed-canopy forests to novel disturbance regimes and suggests that similar settings may be less resilient to climate-induced changes in the future.

human impacts | land cover change | deforestation

The degree to which ecosystems have been altered by human activities is a key criteria for prioritizing conservation efforts globally, particularly in identifying regions that are largely pristine and worthy of protection (1, 2). Paleocological records have become an important tool in such assessments because they disclose the enduring effects of prehistoric forest clearance, cultivation, and pastoralism, as well as the consequences of species introductions and extinctions well before 17th century European expansion (3–8). Globally, ancient land-cover change has been greatest in regions of densely populated agricultural societies (9, 10). The colonization of islands in Eastern Polynesia by agricultural-based societies [from *circa* (*ca.*) A.D. 800] was no exception and was almost always associated with extensive forest clearance, erosion, and biotic extinctions through the introduction of fire, cultigens, weeds, and mammalian predators (7, 11–14). Similarly, Polynesian (Māori) arrival in the South Island of New Zealand 700–800 y ago was followed by clearance of more than 40% of the native forests (12, 15, 16). What is remarkable is that this extensive deforestation was accomplished by small, largely transient, nonagricultural populations in places remote from any settlement, and the forest loss occurred throughout the relatively large South Island (151,215 km<sup>2</sup>) in only a few decades. In many areas the native forest has not recovered (12, 16).

Although the fact that once-extensive forests had vanished was well documented in the 19th century, a long debate ensued as to whether people or climate changes were responsible (15, 17). McGlone (15, 17) concluded that climate change was not a significant factor, based on a synthesis of the relatively few, poorly dated pollen diagrams and subfossil wood-charcoal data that were available. Elsewhere in the world, where people have been present throughout the Holocene, climate changes are strongly implicated in changing fire frequencies (18). It therefore has remained a possibility that Māori settlement of New Zealand coincided with a climate regime unusually suited to large-scale fire. Here, we reexamine the hypothesis that the rapid deforestation of New Zealand was due to human influence alone by examining the climate, fire, and environmental changes of the last 1000 y across gradients of topography and precipitation.

Before Polynesian arrival in New Zealand, 85–90% of the country was heavily forested with low scrub and herbaceous communities occurring above treeline (15). The South Island supported continuous closed-canopy forests dominated by *Nothofagus* spp. (beech) at wetter, higher elevations, and podocarps in drier, lower elevations (predominantly represented by *Dacrydium cupressinum*, *Prumnopitys* spp., *Dacrycarpus dacrydioides*, *Podocarpus* spp., *Halocarpus* spp. and *Phyllocladus alpinus*), as evidenced by a network of pollen records (12). A low podocarp-dominated scrub with some grassland was widespread below treeline in the semiarid center of Otago and South Canterbury (12), but otherwise, extensive scrub was rare. Māori arrived in New Zealand *ca.* A.D. 1280 (19, 20), probably in a founding group that included at least 50–100 women (21). Across most of the South Island, charcoal preserved in soils suggests that fire occurrence prior to human arrival was rare (1–2 events/millennia) (22) and, with few exceptions (23), ecologically insignificant. By the time of European settlement in the mid 19th century, over 40% of the island's forests had been cleared by fire (15, 24) and replaced by tussock grassland and fern-shrubland (25–27). Subsequent forest clearance by Europeans transformed the remaining fern-shrubland to pastureland with nonnative plants.

Recent studies have documented the vegetation changes (12) and refined the timing of human arrival (19), but the relative importance of anthropogenic burning and climate change in driving the deforestation, and the character of fire events and ecological and limnobiologic responses, have not been addressed (16). To better understand the sequence of events that led to the demise of

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extensive native forests in the South Island of New Zealand, we reconstruct the watershed history of 16 sites, based on paleoenvironmental data preserved in the sediments of 16 small closed-basin lakes that span a range of environmental conditions (Fig. 1 and Table S1), from low-elevation watersheds [0–300 m above sea level (masl)], once dominated by podocarp forests, to cooler, moister middle-elevation (300–600 masl), and wet watersheds at high elevations [ $>600$  masl,  $>1600$  mm mean annual precipitation (ppt)] that formerly supported closed-canopy beech forest. For each site, we derived chronologies from accelerator mass spectrometry (AMS) radiocarbon-dated sediment cores (Table S2) and analyzed macroscopic charcoal data to identify individual fire events (charcoal peaks representing one or more fires during the time span of the peak) and patterns and trends in overall burning. Pollen records revealed the sequence of vegetation changes, geochemistry and magnetic susceptibility data indicated changes in erosion and slope stability, and diatom and chironomid data showed changes in the limnobiota.

### Results and Discussion

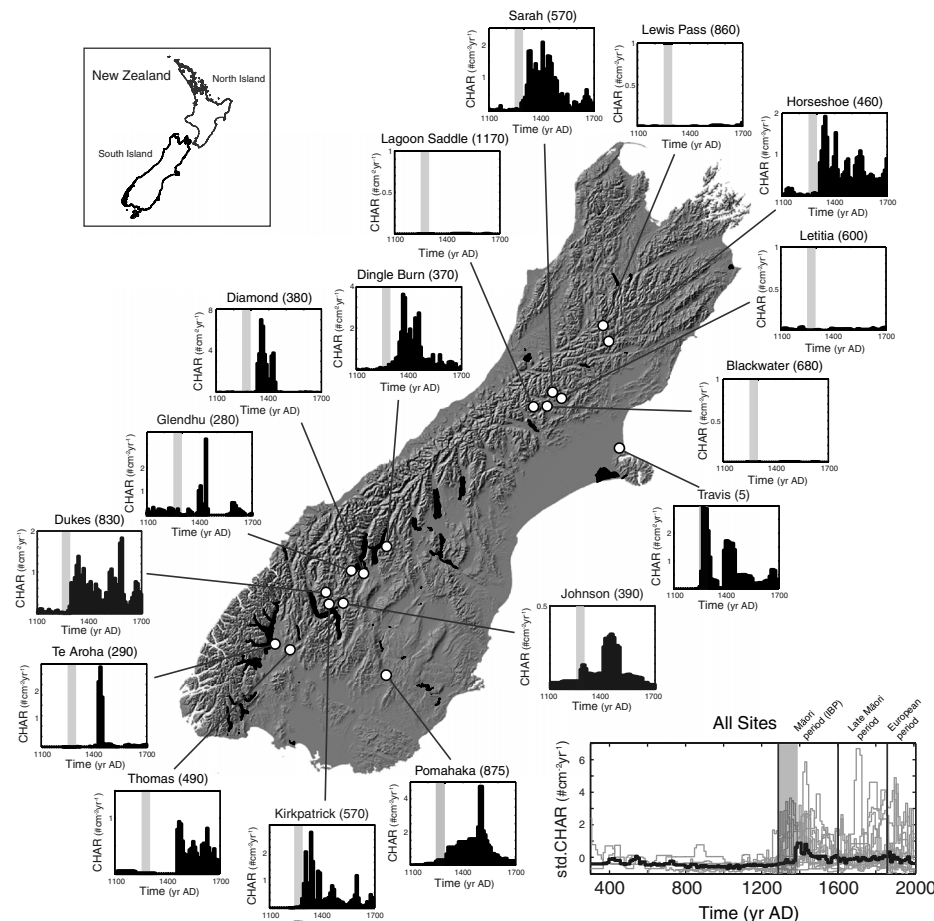
Within 200 y of Māori arrival, charcoal records from all but the wettest ( $>1600$  mm ppt), high-elevation ( $>800$  masl) locations registered a dramatic increase in macroscopic charcoal abundance, suggesting a period of intense local fire activity (Fig. 1). This initial burning period (IBP) (16) consisted of one to three fire events per century in individual watersheds. While the IBP was not synchronous across the South Island (within dating error), it occurred between ca. A.D. 1280–1600, except at sites with high rainfall and where geographic barriers (e.g., large lakes or rivers) likely prevented fire spread. Local fires during the IBP were large and/or severe (Fig. 1), judging from the magnitude of individual charcoal peaks: Mean peak magnitude during the IBP was  $48.17$  particles  $\text{cm}^{-2} \text{y}^{-1}$  (95% C.I.s = 20.09 to 76.26),

whereas mean peak magnitude prior to ca. A.D. 1280 was 2.13 (C.I.s =  $-1.61$  to 5.86).

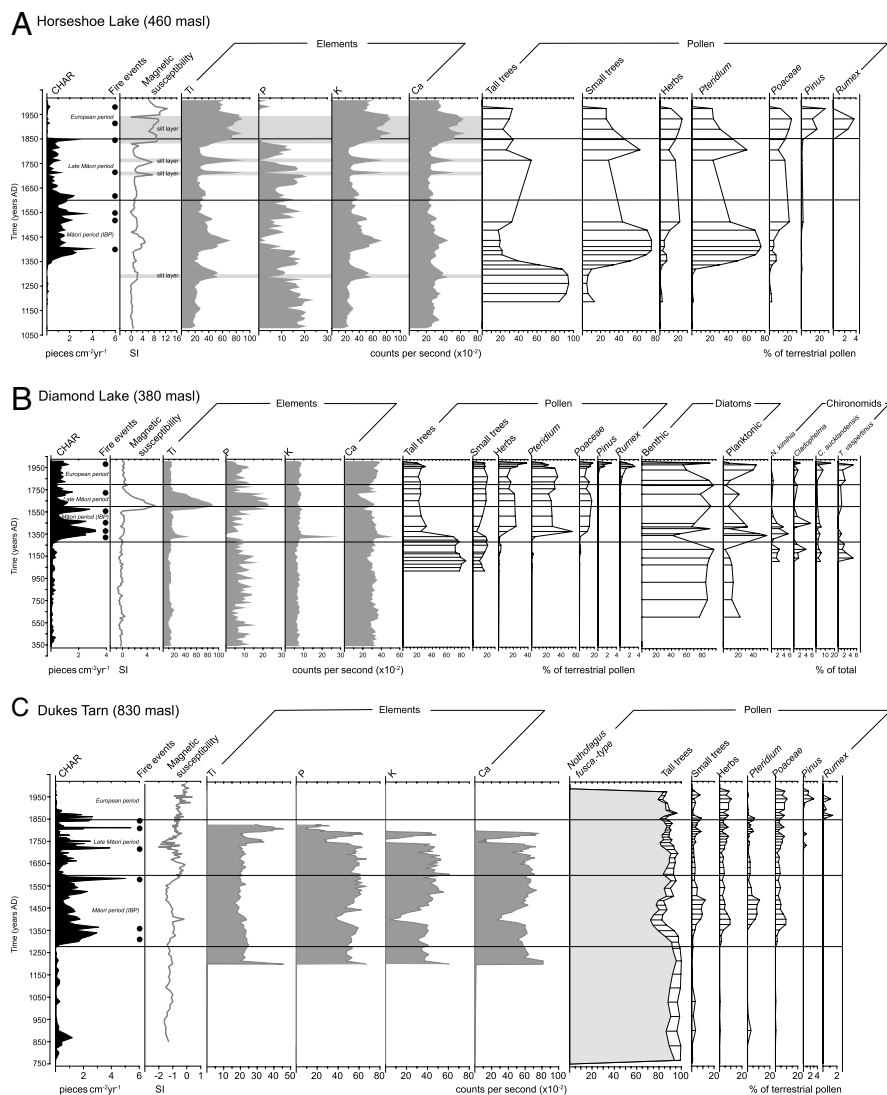
Dendroclimatological and speleothem data registered considerable climate variability during the last 1000 y (28–30), but climate changes were not associated with the fires of the IBP. Superposed epoch analysis (SEA) (31, 32) allowed comparison of charcoal levels and fire events with extreme summer temperatures; the latter were reconstructed from silver pine (*Lagarostrobos colensoi*) tree-ring chronologies and calibrated with instrumental data ( $>95\%$  C.I. of mean summer temperatures from ca. A.D. 850–2000). The results showed no statistical relationship between anomalous summer temperatures (Fig. S1) (29, 30) and the IBP at particular sites, confirming an anthropogenic origin for the fires.

The charcoal records from most sites showed a pattern of several high-magnitude fire events (charcoal peaks) during the decades that define the IBP followed by a period of less-frequent and less-severe fire events during the Late Māori period ( $\sim 1600$ –1840 A.D.) and a second increase in the frequency and intensity of fire activity following European arrival (A.D. 1840 to present) (Fig. 1, Bottom Right). A few sites experienced sustained burning from the onset of human arrival until present (e.g., Lake Kirkpatrick and Travis Swamp), and several sites showed little evidence of fire occurrence during the past 1000 y (e.g., Lewis Pass, Lagoon Saddle, and Blackwater Lake).

Sites where pollen data were also available (Horseshoe Lake, Diamond Lake, Dukes Tarn, Travis Swamp, Pomahaka Bog, and Lake Thomas) indicate that the most severe deforestation during the IBP occurred in moderate or low rainfall regions ( $<1600$  mm ppt), where large fire events caused a shift from beech and podocarps to bracken (*Pteridium esculentum*), grasses, and small trees and shrubs (e.g., *Coprosma* spp., *Coriaria* spp., and *Kunzea ericoides*) (Fig. 2). Sites exhibiting some forest recovery in the



**Fig. 1.** Charcoal accumulation rates (CHAR pieces  $\text{cm}^{-2} \text{y}^{-1}$ ) and location of 16 sites, South Island, New Zealand. Site names above plots and elevations (masl). Small plots show variation in CHAR between A.D. 1100–1700. (Bottom Right) Composite plot with standardized CHAR levels for each site (gray lines) and mean CHAR for all 16 sites (thick black line) for the time period A.D. 300–2000. Gray band in both small plots and large plot indicates estimated time of Polynesian (Māori) arrival in New Zealand ca. A.D. 1280 (19).



**Fig. 2.** Paleoenvironmental data for sites with varying degrees of human impact, South Island New Zealand. Variation in CHAR, pollen percentages, magnetic susceptibility, and element concentrations for (A) Horseshoe Lake, (B) Diamond Lake, and (C) Dukes Tarn. Diamond Lake plot (B) also shows changes in diatom and chironomid community composition.

Late Māori period lie in moderately wet high-elevation settings that currently experience no measurable annual water deficits (e.g., Dukes Tarn and Pomahaka Bog, 1000–2000 mm; Table S1). Sites with no fire activity in the last 1000 y are located in wet (>1600 mm ppt), high-elevation (>800 masl) settings (e.g., Lewis Pass) or protected locations (e.g., Lake Letitia, which was likely spared from fire spread by the broad Waimakariri River floodplain). Dukes Tarn, a moderately wet (~1340 mm ppt) high-elevation (830 masl) site, recorded persistent fire activity in the IBP and Late Māori period, but only modest impacts on vegetation.

Changes in patterns of fire, vegetation, erosion, slope stability, and lake chemistry were more fully examined at three lakes (Horseshoe Lake, Diamond Lake, and Dukes Tarn), and limnobiological responses were analyzed at Diamond Lake. These lakes represent three different settings: a dry, middle-elevation (460 masl) site situated along what was known to be an important east–west travel way (Horseshoe Lake); a dry (980 mm ppt), middle-elevation (380 masl) site adjacent to Lake Wanaka (Diamond Lake); and a moderately wet (>1342 mm ppt) high-elevation (830 masl) site near treeline in the Southern Alps (Dukes Tarn).

**Horseshoe Lake (Latitude 42.60° South, Longitude 172.52° East; 460 masl).** Horseshoe Lake is a middle-elevation site located on the south side of Lewis Pass in the central foothills of the South Island and lies along the seismically active Hope Fault (33) (Fig. 1). Charcoal and pollen data from Horseshoe Lake suggest that the IBP began *ca.* A.D. 1370 and involved three large-mag-

nitude fire events (Fig. 2A). The pollen data record a rapid decline in closed-canopy forest taxa (from 90–20% of terrestrial pollen taxa) and a rise in ferns (from 5–75%) and grassland (from 5–20%; Fig. 2A). Peaks in magnetic susceptibility and element concentrations (titanium, potassium, and calcium) match silt layers that date to known seismic activity along the Hope Fault as well as to increased fire activity; at the same time, phosphorus levels drop precipitously. Prior to the IBP, increases in magnetic susceptibility and titanium, potassium, and calcium match a layer of silt that coincides with an earthquake in A.D. 1316 (33). Magnetic susceptibility and titanium, potassium, and calcium also peaked immediately following large-magnitude IBP fire events (versus steep declines in phosphorus), especially *ca.* A.D. 1400–1450 and suggest that increased fire activity led to soil instability and large-magnitude erosion.

During the Late Māori period, *ca.* A.D. 1600–1840, three fire events occurred *ca.* A.D. 1610, 1720, and 1840. Episodic but less-frequent fire events (<1 per century) were associated with an increase in *Nothofagus* and Podocarpaceae pollen, suggesting partial recovery of native forest. Brief (years to decades) but dramatic spikes in magnetic susceptibility and titanium, potassium, and calcium coincide with two silt layers *ca.* A.D. 1720 and 1770; the first corresponds with a fire event, and the second may be associated with a seismic event in the mid-18th century (33).

In the European period, two small fire events (*ca.* A.D. 1920 and 1980) are matched by a pollen increase from exotic plants (*Pinus* and *Rumex acetosella*) (Fig. 2A). Sustained increases in

magnetic susceptibility, titanium, potassium, and calcium at the same time that phosphorus dropped precipitously to near undetectable levels (*ca.* A.D. 1840–1960) correspond to a wide band of silt in the core (Fig. 2A.), most likely a result of earthquake destabilized slopes from a large-magnitude earthquake in A.D. 1888 (33) and possibly also from steady leaching of deforested substrates following fires. Repeated burning of seral vegetation probably limited forest recovery. Such surface fires are not easily detected in the macroscopic charcoal record (34) but may be inferred from low-magnitude charcoal accumulation rates (CHAR) and the steady loss of phosphorus.

The multiple proxy records from Horseshoe Lake portray a history of earthquakes, fires, and changing land use. These records indicate that significant burning in the IBP led to a shift from closed-canopy forest to a fern-shrubland, increased erosion, and steady nutrient transfer from the watershed to the lake. The Late Māori period and European periods were characterized by small fire events marked by peaks in erosion, highly variable nutrient levels, and little forest recovery. The magnitude of European land-use change at Horseshoe Lake is similar to that observed at sites in the North Island of New Zealand, where sheep farming and slope instability have been intense (35).

**Diamond Lake (Latitude 44.65° South, Longitude 168.96° East; 380 masl).** Diamond Lake is a middle-elevation site located 2 km from the southern shore of Lake Wanaka (Fig. 1). Charcoal data suggest four large-magnitude fire events occurred during the IBP at *ca.* A.D. 1320, 1340, 1460, and 1560 (Fig. 2B). These events resulted in a decline in forest taxa (pollen percentages decline from 80 to 20%) and an increase in ferns (from <5–60%) and grasses (from <5–15%) (Fig. 2B). Large peaks in magnetic susceptibility and element concentrations (titanium, potassium, and calcium) at the beginning and end of the IBP (*ca.* A.D. 1330 and 1590) suggest substantial erosion in the watershed, following fire events in A.D. 1320 and 1560. Phosphorus levels in Diamond Lake decreased to low levels following the first indication of fire activity during the IBP (*ca.* A.D. 1330) but then increased substantially in the late 16th century. The IBP was also associated with a shift in diatom and chironomid communities. For example, prior to the IBP the diatom community was dominated by benthic and tychoplanktonic diatom species that tolerated a large range of nutrient conditions and would have inhabited large areas of the lake bottom. After the IBP, planktonic species dominated the assemblages, especially *Discostella stelligera*, which has been shown to respond positively to phosphorus additions (36) and would have benefited from inputs of silica from erosional events (Fig. 2B). Increased terrestrial inputs could have also reduced light availability to the lake bottom, reducing the area of benthic habitat. Likewise, chironomid species assemblages indicate a shift from *ca.* A.D. 1300–1430 (Fig. 2B). Prior to humans, *Naonella kimihia* was generally low, and abundances of species associated with low nutrient levels and benthic habitat (e.g., *Cladopelma* and *Tanytarsus vespertinus*) were high. During the IBP, *Naonella kimihia*, which tolerates relatively high conductivity that often result from terrestrial inputs (nutrients and dissolved solids), peaked with an erosion event *ca.* A.D. 1320, and *Cladopelma* and *Tanytarsus vespertinus* decreased.

Fire activity was low during the Late Māori period, with only a single small fire event *ca.* A.D. 1750. The watershed supported a mixed fern-shrubland with grasses, and a period of erosion occurred from *ca.* A.D. 1600–1700 (as evidenced by high magnetic susceptibility and titanium levels). An interval between fires *ca.* A.D. 1340–1460 and a decrease in fire activity after *ca.* A.D. 1560 are associated with the return of benthic diatom species (e.g., *Staurosirella pinnata* and *Staurosira* spp.) as well as chironomid species *Cladopelma*, which dominated prior to human arrival.

Renewed fires and deforestation in the European period were associated with an increase in grasses (often sown after deliberate

fires) and the establishment of nonnative plant taxa. Novel diatom communities emerged, as evidenced by increases in eulanktonic diatom taxa *Fragilaria tenera* and *Asterionella formosa*, which are strong competitors for phosphorus and nitrogen and have moderate to high silica requirements (37, 38). Nutrient leaching from land-use activities and the introduction of *A. formosa* may have driven this diatom shift (39). Similarly, the chironomid assemblage during the last 100 y is quite different from any previous time, with increasing numbers of *Cricotopus aucklandensis*, a species tolerant of high conductivity and nutrient inputs associated with erosional events.

Paleoenvironmental data from Diamond Lake suggests that high-magnitude fire events during the IBP altered vegetation and watershed conditions, and the ecological response included dynamic species turnover of the limnobiota. During the Late Māori period, intervals of decreased fire activity or lower-magnitude fires (as evidenced by peak magnitude) suggest some level of postfire recovery of vegetation as well as diatom and chironomid community composition to a near pre-Māori state. European burning marked a second shift in vegetation, the introduction of nonnative plants and the establishment of novel diatom and chironomid assemblages.

**Dukes Tarn (Latitude 44.96° South, Longitude 168.49° East; 830 masl).** Dukes Tarn is located 2 km east of the eastern shores of Lake Wakatipu at relatively high elevation (830 masl) (Fig. 1). Charcoal data register fire events during the IBP at *ca.* A.D. 1290, 1360, and 1570, and the pollen data indicate that forests recovered within decades (Fig. 2C). For example, following evidence of fire activity in *ca.* A.D. 1360, percentages of *Pteridium* and *Poaceae* rose from <5% to >10% and fell to low levels (<5%) at *ca.* A.D. 1450 (Fig. 2C). Phosphorus, calcium, and potassium levels declined abruptly at *ca.* A.D. 1380 and *ca.* A.D. 1750, followed by a return to pre-IBP levels. Fire activity continued into the Late Māori period; native forest taxa declined and open vegetation increased modestly *ca.* A.D. 1750–1850. No large fire events occurred in the European period, and introduced plant taxa mark conversion to pastureland in the mid-19th century. Unlike the other two sites, there was some recovery of native vegetation within decades of the IBP and European burning periods, and fires seem to have had little lasting impact on the long-term structure and composition of vegetation.

## Conclusions

Our results confirm that human-set fires were responsible for the loss of New Zealand's forests, and that the severity and consequences of the fires varied with geography and local climate. Interannual to decadal climate variations recorded in tree-ring data do not suggest that dry or warm periods coincided with the IBP, although we cannot rule out seasonal climate variability as a possible determinant of fire size and severity. Among our 16 watersheds, sites at low and middle elevations (<0–600 masl) and dry settings (<1600 mm ppt) showed wholesale conversion of closed forest to tussock grasslands and fern-shrubland and little subsequent recovery. Higher-elevation watersheds in wet settings (>1600 mm ppt) registered substantially less impact during the IBP, and pollen data from one high-elevation site shows forest recovery during the Late Māori period (Fig. 2C). The fire activity during the IBP was also determined by site location. Those sites isolated from travel ways (40, 41) or surrounded by natural fire breaks also experienced fewer fire events and less deforestation. Drier, lower-elevation sites with larger and more permanent human settlement (e.g., Travis Swamp; <650 mm ppt) had persistent fires from the onset of Māori arrival to the present (42).

Fires initiated forest destruction, slope instability, and erosion. The import of sediments and nutrients evoked a response in the lake chemistry and limnobiota, as evidenced by shifts in diatom and chironomid community composition. In some locations, forest and limnobiota recovered in the Late Māori period to their pre-

IBP conditions within a few decades, but sites in the interior region remained open landscapes, probably from sustained burning. European arrival led to increased fires and the conversion of open tussock grasslands and fern-shrublands to pastureland. In a few watersheds, European impacts via erosion, nutrient transport, and eutrophication were more severe than in the IBP (35, 43).

When compared with other islands in the Pacific, and even the North Island of New Zealand, the rate and scale of deforestation across a topographically diverse landscape is impressive. How then did so few people manage to transform so much of the South Island landscape within 200 y? Polynesians had a long history of island colonization that informed their use and management of New Zealand landscapes (7, 11, 13, 44–46). Successful establishment of Polynesian populations typically involved intensive use and management of wild terrestrial, freshwater, and marine resources [for example, in New Zealand, moa (family Dinornithidae), freshwater eels (family Anguillidae), and cockles (*Austrovenus stutchburyi*)]. There is no evidence that New Zealand forests were cleared to facilitate the hunting of birds; rather, birds were snared or speared along forest tracks. Management of natural resources, the cultivation of important food crops, and the harvesting of edible wild foods required efficient methods of land clearance and subsequent stewardship to maintain open landscapes and varied vegetation (15). In the case of the South Island, archaeological evidence suggests that successful cultivation of introduced food crops such as kumara (*Ipomoea batatas*) and taro (*Colocasia esculenta*) was not possible in the montane interior nor along the coast south of Banks Peninsula (latitude 43° south) (47). In these regions, the rhizomes of postfire-induced bracken and other starch-rich plants, such as *Cordyline australis*, were an important source of carbohydrates in Māori diets (27, 48). Māori, in their efforts to increase the productivity of beech and podocarp forests for their uses, encouraged more heterogeneous (both structurally and compositionally) and economically useful fern-shrubland at the same time as making travel easier (15, 17, 27). Europeans embarked on a similar campaign to enhance and augment the resource base in the interior South Island by grazing open tussock grasslands and converting shrublands to pasture (45).

The vulnerability of South Island forests to human-set fires was closely linked with a strong west–east rainfall gradient, and forests in areas with high levels of rainfall (>1600 mm ppt) were less impacted by fires than drier eastside forests (15). The wet, cool conditions of the late Holocene limited natural ignitions throughout most of the South Island (28) and likely facilitated the establishment of tree species that were poorly adapted to fire, in that they have thin bark, are poorly adapted to resprouting (e.g., lignotubers), and have seeds that are particularly vulnerable to fire and have relatively short-lived viability (20, 49–51). Even where fire activity decreased after the IBP, recovery of native forests was slow (i.e., centuries), further suggesting the vulnerability of forest species to repeated fire. The clearance by fire of vast tracts of native forests in the drier eastern regions of the South Island illustrates the positive feedbacks that are initiated with the introduction of a new disturbance.

A shift to novel plant communities and nonforested vegetation in New Zealand (*sensu ref.* 52) led to long-term changes in watershed and aquatic characteristics that persist to the present-day in some settings. High-resolution and well-dated paleoecological records from the South Island of New Zealand help to disentangle the influence of humans and climate on vegetation and watershed change since the time of settlement and offer insights into the natural role of fire that could not be inferred by examination of the present-day landscape alone. Historical insights, such as these, provide important information for developing current fire-management plans and conservation strategies.

## Materials and Methods

Age determinations and core chronologies were based on AMS <sup>14</sup>C dates obtained on twig charcoal and terrestrial plant macrofossils (twigs, leaves, and plant fragments) (Table S2). Pollen analysis followed the preparation methods of Moore et al. (53). Pollen counts were continued until a dryland pollen sum of at least 250 grains was reached. Percentage calculations are based on a terrestrial pollen sum (excluding tree fern spores because they tend to be highly overrepresented). The first evidence of exotic *Pinus* (pine) pollen marks the establishment of large pine plantations in the late European period (post-1960s) (54). Pollen types were grouped as closed forest taxa (predominantly *Nothofagus* spp. and *Podocarps*); small trees and seral shrubs and ferns (for example, *Myrsine*, *Coprosma* spp., *Coriaria* spp., *Leptospermum*, *Pseudopanax*, *Rubus*, and *Pteridium esculentum*), herbs (mostly Poaceae and Cyperaceae), and exotic taxa, including *Pinus* spp. and *Rumex acetosella*. These groupings were used to identify the transition from closed forest to fern-shrubland and grassland and European arrival.

High-resolution charcoal analysis followed methods of Whitlock et al. (55). Charcoal particles (>125 μm) were examined to reconstruct local fire events (55) based on changes in CHAR (particles cm<sup>-2</sup> y<sup>-1</sup>). Decomposition of the charcoal time series employed a Gaussian mixture model to identify the mean and variance of the background CHAR distribution (56). The 99th percentile of this distribution is defined as the threshold value separating peak fire events from “noise.” Significant charcoal peaks, determined by CHAR values greater than a locally defined threshold value, identified specific fire events. Because each 1-cm-long core interval in our lakes represents 5–10 y, a charcoal peak may represent one or more fires occurring within the time span of a charcoal peak. The statistical significance of each peak was evaluated by comparing the original charcoal counts against the values in samples occurring 35 y 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 y, then a “peak” was not identified (57). Fire size and/or intensity are inferred from the magnitude of individual charcoal peaks (particles cm<sup>-2</sup> y<sup>-1</sup>) (58).

Magnetic susceptibility was performed at 1-cm-long core intervals with a Geotek XYZ multisection automated split core logger that measured inorganic allochthonous sediment (59). Major, minor, and trace element concentrations were measured to examine changes in nutrients and to reconstruct lake chemistry and its relationship to soil fertility. Mineralogical and geochemical analyses at 1-cm intervals (~5–10 y resolution) were derived from X-ray diffraction mineralogy using an Itrax X-ray fluorescence instrument, and each measurement was taken over an area of 1 cm<sup>2</sup> using a 30-s count time (60, 61).

Slides for diatom identification were prepared by digesting ~100 mg of sediment in 10% HCl and 30% H<sub>2</sub>O<sub>2</sub> to oxidize carbonates and labile organic matter, respectively. Samples were allowed to settle passively, the supernatant was aspirated, and the sample was rinsed three times with deionized water. Diluted slurries were permanently mounted using Naphrax® (62). Diatoms were identified and enumerated with differential interface contrast light microscopy at 1000×. The published floras of Krammer et al. (63–66) were used for identification.

Chironomid head capsules were extracted from 0.5-mL sediment samples following the procedure described by Walker et al. (67). This volume was sufficient to recover 62–96 full head capsules per level. Heads were identified with reference to Dieffenbacher-Krall et al. (68). Cluster analysis of chironomid data was performed by incremental sum-of-squares using the CONISS program (69) to identify periods of major assemblage transition. Canonical correspondence analysis of chironomid and environmental data from Dieffenbacher-Krall et al. (70) surface samples from South Island lakes was performed with Diamond Lake samples to identify potential environmental factors associated with assemblage changes.

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1. Brooks TM, et al. (2006) Global biodiversity conservation priorities. *Science* 313(5783): 58–61.

2. Mittermeier RA, et al. (2003) Wilderness and biodiversity conservation. *Proc Natl Acad Sci USA* 100(18):10309–10313.

3. Willis KJ, Gillson L, Brncic TM (2004) How "virgin" is virgin rainforest? *Science* 304(5669):402–403.
4. Heckenberger MJ, Christian Russell J, Toney JR, Schmidt MJ (2007) The legacy of cultural landscapes in the Brazilian Amazon: Implications for biodiversity. *Philos Trans R Soc B* 362(1478):197–208.
5. Lynch AH, et al. (2007) Using the paleorecord to evaluate climate and fire interactions in Australia. *Annu Rev Earth Planet Sci* 35:215–239.
6. Haberle SG, David B (2004) Climates of change: Human dimensions of Holocene environmental change in low latitudes of the PEPIL transect. *Quat Int* 118–119:165–179.
7. Bayliss-Smith T, Hviding E, Whitmore T (2003) Rainforest composition and histories of human disturbance in Solomon Islands. *AMBIO* 32(5):346–352.
8. Marlon J, et al. (2008) Climate and human influences on global biomass burning over the past two millennia. *Nat Geosci* 1:697–701.
9. Kaplan JO, Krumhardt KM, Zimmerman N (2009) The prehistorical and preindustrial deforestation of Europe. *Quat Sci Rev* 10.1016/j.quascirev.2009.1009.1028.
10. Pongratz J, Reick C, Raddatz T, Claussen M (2008) A reconstruction of global agricultural areas and land cover for the last millennium. *Global Biogeochem Cycles* 22(3):GB3018.
11. Kirch P, Kahn J (2007) Advances in Polynesian prehistory: A review and assessment of the past decade (1993–2004). *J Archaeol Res* 15(3):191–238.
12. McGlone MS, Wilmshurst JM (1999) Dating initial Māori environmental impact in New Zealand. *Quat Int* 59:5–16.
13. Steadman DW (1995) Prehistoric extinctions of Pacific Island birds: Biodiversity meets zooarchaeology. *Science*(267):1123–1130.
14. Rolett BV, Diamond J (2004) Environmental predictors of pre-European deforestation on Pacific Islands. *Nature* 431:443–446.
15. McGlone M (1983) Polynesian deforestation of New Zealand: A preliminary synthesis. *Archaeol Ocean* 18:11–25.
16. McWethy DB, Whitlock C, Wilmshurst JM, McGlone MS, Li X (2009) Rapid deforestation of South Island, New Zealand, by early Polynesian fires. *Holocene* 19(6):883–897.
17. McGlone MS (1989) The Polynesian settlement of New Zealand in relation to environmental and biotic changes. *NZ J Ecol* 12:115–129.
18. Whitlock C, Higuera PE, McWethy DB, Briles CE (2010) Paleoecological perspective on fire ecology: Revisiting the fire regime concept. *Open Ecol J* 3:6–23.
19. Wilmshurst JM, Anderson AJ, Higham TFG, Worthy TH (2008) Dating the late prehistoric dispersal of Polynesians to New Zealand using the commensal Pacific rat. *Proc Natl Acad Sci USA* 105(22):7676–7680.
20. Higham TFG, Anderson AJ, Jacomb C (1999) Dating the first New Zealanders: The chronology of Wairau Bar. *Antiquity* 73:420–427.
21. Murray-McIntosh RP, Scrimshaw BJ, Hatfield PJ, Penny D (1998) Testing migration patterns and estimating founding population size in Polynesia by using human mtDNA sequences. *Proc Natl Acad Sci USA* 95(15):9047–9052.
22. Ogden J, Basher L, McGlone M (1998) Fire, forest regeneration and links with early human habitation: Evidence from New Zealand. *Ann Bot* 81(6):687–696.
23. McGlone MS, Moar NT (1998) Dryland Holocene vegetation history, Central Otago and the Mackenzie Basin, South Island, New Zealand. *NZ J Bot* 36(1):91–111.
24. Cumberland K (1962) "Climate change" or cultural interference? New Zealand in Moahunter times. *Land and Livelihood*, ed M McCaskill (NZ Geogr Soc, Christchurch, New Zealand), pp 88–142.
25. McGlone MS (2001) The origin of the indigenous grasslands of southeastern South Island in relation to pre-human woody ecosystems. *NZ J Ecol* 25(1):1–15.
26. Mark A, McLennan B (2005) The conservation status of New Zealand's indigenous grasslands. *NZ J Bot* 43:245–270.
27. McGlone MS, Wilmshurst JM, Leach HM (2005) An ecological and historical review of bracken (*Pteridium esculentum*) in New Zealand, and its cultural significance. *NZ J Ecol* 29(2):165–184.
28. Lorrey A, et al. (2008) Speleothem stable isotope records interpreted within a multiproxy framework and implications for New Zealand palaeoclimate reconstruction. *Quat Int* 187(1):52–75.
29. Cook ER, Palmer JG, D'Arrigo RD (2002) Evidence for a "Medieval Warm Period" in a 1,100 year tree-ring reconstruction of past austral summer temperatures in New Zealand. *Geophys Res Lett* 10.1029/2001GL014580.
30. Cook K, Visky E (2006) South American climate during the last glacial maximum: Delayed onset of the South American monsoon. *J Geophys Res—Atm* 3:1–21.
31. Grissino-Mayer HD (1995) Tree-ring reconstructions of climate and fire history at El Malpais National Monument, New Mexico. Dissertation (Univ of Arizona, Tucson, AZ).
32. Grissino-Mayer HD, Swetnam TW (2000) Century-scale climate forcing of fire regimes in the American Southwest. *Holocene* 10(2):213–220.
33. Cowan HA, McGlone MS (1991) Late Holocene displacements and characteristic earthquakes on the Hope River segment of the Hope Fault, New Zealand. *J Roy Soc NZ* 21(4):373–384.
34. Whitlock C, Skinner CN, Bartlein PJ, Minckley T, Mohr JA (2004) Comparison of charcoal and tree-ring records of recent fires in the eastern Klamath Mountains, California, USA. *Can J Forest Res* 34:2110–2121.
35. Wilmshurst JM (1997) The impact of human settlement on vegetation and soil stability in Hawke's Bay, New Zealand. *NZ J Bot* 35:97–111.
36. Anderson NJ, Blomqvist P, Renberg I (1997) An experimental and palaeoecological study of algal responses to lake acidification and liming in three central Swedish lakes. *Eur J Phycol* 32:35–48.
37. Interlandi SJ, Kilham SS, Theriot EC (1999) Responses of phytoplankton to varied resource availability in large lakes of the greater Yellowstone ecosystem. *Limnol Oceanogr* 44:668–682.
38. Saros JE, Michel TJ, Interlandi SJ, Wolfe AP (2005) Resource requirements of *Asterionella formosa* and *Fragilaria crotonensis* in oligotrophic alpine lakes: Implications for recent phytoplankton community reorganizations. *Can J Fish Aquat Sci* 62:1681–1689.
39. Harper MAKocielek JP (1990) Did Europeans introduce *Asterionella Formosa* Hassall to New Zealand? *Proceedings of the 11th International Diatom Symposium* (California Academy of Sciences, San Francisco), pp 479–484.
40. Gardner WJ (1983) *The Amuri, a Country History* (Caxton Press, Christchurch).
41. Brailsford B (1996) *Greenstone Trails: The Māori and Pounamu* (Stoneprint, Hamilton), 2nd ed.
42. Rose RA (2004) Holocene fire regimes reconstructed from peat core charcoal analysis in the South Island, New Zealand. Masters Thesis (University of Canterbury, Christchurch, NZ).
43. Glade T (2003) Landslide occurrence as a response to land use change: A review of evidence from New Zealand. *CATENA* 51(3–4):297–314.
44. Anderson A (2002) Faunal collapse, landscape change and settlement history in remote Oceania. *World Archaeol* 33:375–390.
45. Pawson E, Brooking T (2002) *Environmental Histories of New Zealand* (Oxford Univ Press, Melbourne).
46. Anderson AJ, Smith IWG (1996) The transient village in southern New Zealand. *World Archaeol* 27:359–371.
47. Furey L (2006) *Māori Gardening: An Archeological Perspective* (New Zealand Department of Conservation, Wellington, New Zealand) p 137.
48. Hamel J (2001) *The Archaeology of Otago* (New Zealand Department of Conservation, Wellington, New Zealand).
49. Baylis G (1980) Mycorrhizas and the spread of beech. *NZ J Ecol* 3:151–153.
50. Wardle P (1980) Ecology and distribution of silver beech (*Nothofagus menziesii*) in the Paringa District, South Westland, New Zealand. *NZ J Ecol* 3:23–36.
51. Wardle P (1984) *The New Zealand Beeches: Ecology, Utilization, and Management* (New Zealand Forest Service, Christchurch, NZ).
52. Williams JW, Jackson ST (2007) Novel climates, no-analog communities, and ecological surprises: Past and future. *Front Ecol Environ* 5:475–482.
53. Moore PD, Webb JA, Collinson ME (1991) *Pollen Analysis* (Blackwell Scientific, Oxford, UK).
54. Roche M (1990) *History of New Zealand Forestry* (New Zealand Forestry Corporation in association with GP Books, Wellington, NZ) p 53.
55. Whitlock C, Larsen C (2001) Charcoal as a Fire Proxy. *Tracking Environmental Change Using Lake Sediments: Volume 3 Terrestrial, Algal, and Siliceous Indicators*, eds JP Smol, HJB Birks, and WM Last (Kluwer Academic Publishers, Dordrecht, The Netherlands), pp 75–97.
56. Higuera PE, Peters ME, Brubaker LB, Gavin DG (2007) Understanding the origin and analysis of sediment-charcoal records with a simulation model. *Quat Sci Rev* 26(13–14):1790–1809.
57. Higuera PE, Brubaker LB, Anderson PM, Hu FS, Brown TA (2009) Vegetation mediated the impacts of postglacial climate change on fire regimes in the south-central Brooks Range, Alaska. *Ecol Monogr* 79(2):201–219.
58. Whitlock C, Brunelle A (2006) Pollen records from northwestern North America. *Enc Quat Sci*, ed S Elias 1170–1178.
59. Gedye SJ, Jones RT, Tinner W, Ammann B, Oldfield F (2000) The use of mineral magnetism in the reconstruction of fire history: A case study from Lago di Origlio, Swiss Alps. *Palaeogeogr Palaeoclimatol* 164(1–4):101–110.
60. Kuhlmann H, Freudenthal T, Helmke P, Meggers H (2004) Reconstruction of paleoceanography off NW Africa during the last 40,000 years: Influence of local and regional factors on sediment accumulation. *Mar Geol* 207(1–4):209–224.
61. Jansen JHF, Gaast SJVD, Koster B, Vaars AJ (1998) CORTEX, a shipboard XRF-scanner for element analyses in split sediment cores. *Mar Geol* 151(1–4):143–153.
62. Battarbee R, et al. (2001) Diatoms. *Tracking Environmental Change Using Lake Sediments: Volume 3 Terrestrial, Algal, and Siliceous Indicators*, eds JP Smol, HJB Birks, and WM Last (Kluwer Academic Publishers, Dordrecht, The Netherlands), pp 155–202.
63. Krammer K, Lange-Bertalot H ( ) Bacillariophyceae, 1. Teil: Naviculaceae. *Süßwasserflora von Mitteleuropa, Band 2/1*, eds H Ettl, G Gärtner, J Gerloff, H Heynig, and D and Mollenhauer (Gustav Fischer Verlag, Stuttgart/New York) p 876.
64. Krammer K, Lange-Bertalot H (1988) Bacillariophyceae, 2. Teil: Bacillariaceae, Epithemiaceae, Surirellaceae. *Süßwasserflora von Mitteleuropa, Band 2/2*, eds H Ettl, G Gärtner, J Gerloff, H Heynig, and D Mollenhauer (Gustav Fischer Verlag, Stuttgart/New York) p 596.
65. Krammer K, Lange-Bertalot H (1991) Bacillariophyceae, 3. Teil: Centrales, Fragilariaceae, Eunotiaceae. *Süßwasserflora von Mitteleuropa, Band 2/3*, eds H Ettl, G Gärtner, J Gerloff, H Heynig, and D Mollenhauer (Gustav Fischer Verlag, Stuttgart/Jena) p 576.
66. Krammer K, Lange-Bertalot H (1991) Bacillariophyceae, 4. Teil: Achnanthaceae Kritische Ergänzungen zu Navicula (Lineolatae) and Gomphonema. *Süßwasserflora von Mitteleuropa, Band 2/4*, eds H Ettl, G Gärtner, J Gerloff, H Heynig, and D Mollenhauer (Gustav Fischer Verlag, Stuttgart/Jena) p 437.
67. Walker IR (2001) Midges: Chironomidae and related Diptera. *Tracking Environmental Change Using Lake Sediments, vol. 4: Zoological Indicators*, eds JP Smol, HJB Birks, and WM Last (Kluwer Academic Publishers, Dordrecht, The Netherlands) p 217.
68. Dieffenbacher-Krall AC, Vandergoes MJ, Woodward CA, Boothroyd IKG (2008) *Guide to Identification and Ecology of New Zealand Subfossil Chironomids Found in Lake Sediment* (Climate Change Inst, Univ of Maine, Orono, ME).
69. Grimm EC (1987) CONISS: A FORTRAN 77 program for stratigraphically constrained cluster analysis by the method of incremental sum of squares. *Comput Geosci* 13:13–35.
70. Dieffenbacher-Krall AC, Vandergoes MV, Denton GH (2007) An inference model for mean summer air temperatures in the Southern Alps, New Zealand, using subfossil chironomids. *Quat Sci Rev* 26:2487–2504.