A comparison of historical and paleoseismicity in a newly formed fault zone and a mature fault zone, North Canterbury, New Zealand

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Abstract. The timing of large Holocene prehistoric earthquakes is determined by dated surface ruptures and landslides at the edge of the Australia-Pacific plate boundary zone in North Canterbury, New Zealand. Collectively, these data indicate two large ($M > 7$) earthquakes during the last circa 2500 years, within a newly formed zone of hybrid strike-slip and thrust faulting herein described as the Porter's Pass-to-Amberley Fault Zone (PPAFZ). Two earlier events during the Holocene are also recognized, but the data prior to 2500 years are presumed to be incomplete. A return period of 1300–2000 years between large earthquakes in the PPAFZ is consistent with a late Holocene slip rate of 3–4 mm/yr if each displacement is in the range 4–8 m. Historical seismicity in the PPAFZ is characterized by frequent small and moderate magnitude earthquakes and a seismicity rate that is identical to a region surrounding the structurally mature Hope fault of the Marlborough Fault System farther north. This is despite an order-of-magnitude difference in slip rate between the respective fault zones and considerable differences in the recurrence rate of large earthquakes. The magnitude-frequency distribution in the Hope fault region is in accord with the characteristic earthquake model, whereas the rate of large earthquakes in the PPAFZ is approximated (but over predicted) by the Gutenberg-Richter model. The comparison of these two fault zones demonstrates the importance of the structural maturity of the fault zone in relation to seismicity rates inferred from recent, historical, and paleoseismic data.

Introduction

Earthquake hazard assessment traditionally incorporates the analysis of historic and contemporary seismicity, supplemented where possible by geological information about average slip rates, rupture lengths and slip per event. The timing and magnitude of prehistoric earthquakes resulting in surface rupture are determined by dating features offset by the fault and by examining the geometry of the fault trace. However, these data alone often do not provide sufficient information to characterize large paleoseismic events [e.g., Ward, 1994].

In North Canterbury, New Zealand, we have supplemented data from fault traces with data from landslides distributed throughout the eastern foothills of the Southern Alps. These data are used collectively to assess the late Holocene chronology of large

earthquakes in the Porter's Pass-to-Amberley Fault Zone (PPAFZ), a region of newly formed strike-slip faulting that defines the deformation front at the edge of the Australia-Pacific plate boundary zone in northeastern South Island.

North Canterbury straddles the edge of the Australia-Pacific plate boundary zone between the Hikurangi Trough and the Southern Alps (inset, Figure 1a). Dextral strike-slip faults of the Marlborough Fault System (MFS), farther north, transfer oblique plate motion between the Alpine fault and the Hikurangi subduction zone. These faults have smaller cumulative offsets toward the southeast but higher slip rates that reflect a southward relative migration of the locus of deformation across the region during the late Quaternary [e.g., Yeats and Berryman, 1987; H. Anderson *et al.*, 1993]. The most southerly element of the MFS is the Hope fault (Figure 1a). Sixty kilometers south of the Hope fault, a zone of disseminated shear is evolving along a similar trend (Figure 1a) [Rynn and Scholz, 1978; Carter and Carter, 1982; Herzer and Bradshaw, 1985]. Elements of this zone comprise anastomosing strike-slip and thrust faults that extend east-northeast from the Southern Alps along the northern margin of the Canterbury Plains and pass into a fold and thrust belt near the town of Amberley (Figures 1a and 2). We use the name Porter's Pass-to-Amberley Fault Zone (PPAFZ) to describe the strike and approximate

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Figure 1a. Major structural elements of the northeastern South Island, and the locations and years of selected historical earthquakes. The onshore structure is modified from Gregg [1964], Cowan [1992] and unpublished data of the Active Tectonics Research Group, University of Canterbury. Offshore structure is simplified from Barnes [1994]. The numbered rectangles delineate two regions selected for analysis of historical seismicity. Abbreviations are MFS, Marlborough Fault System; HPF, Hope fault; HRS, Hope River segment; KPF, Kakapo fault; PPAFZ, Porter's Pass-to-Amberley Fault Zone; PBF, Pegasus Bay fault; PP, Porter's Pass; and AY, Amberley. Inset shows location of Figures 1a and 1b in relation to the Alpine Fault (AF) and Hikurangi Trough (HT), the main elements of the New Zealand plate boundary. The arrow indicates the direction and magnitude (42 mm/yr) of the plate motion vector [DeMets et al., 1990].

geographic end-points of the zone, which has only recently been mapped comprehensively [Cowan, 1992].

The region between the Hope fault and PPAFZ is one of transition from subduction to collision [Reyners and Cowan, 1993], where the geological structure of the upper crust is dominated by northeast striking reverse and thrust faults that dip southeast and are closely associated with growing folds [Yousif, 1987; Nicol et al., 1994]. This region has accommodated ~12-15% northwest-southeast shortening during the Pleistocene, and rates of shortening close to the Pacific coast are about 1%/100 kyr [Nicol et al., 1994]. Detailed studies along the edge of the plate boundary zone in North Canterbury indicate that deformation probably commenced within the last 0.5-1 m.y., disrupting a volume of crust not significantly deformed since the Cretaceous [Cowan, 1992; Nicol et al., 1994].

The purpose of this paper is to evaluate the late Holocene record of large earthquakes in the PPAFZ, in order to understand the frequency-magnitude distribution of earthquakes in a newly formed fault zone. We compare historical and paleoseismic data from the PPAFZ with data from the adjacent structurally mature Hope fault zone, and use these data to evaluate differences in fault behavior between structurally mature and newly formed fault zones.

Historic Seismicity of the North Canterbury Region

Earthquakes within and adjoining the North Canterbury region during the period 1942-1994 are shown in Figure 1b, with those greater than magnitude 6.5 since 1888 indicated on Figure 1a. The focal depths for most events in the catalogue are uncertain due to the wide spacing of national network seismographs (typically >50 km), but local studies of microseismicity, recorded on portable, small-aperture seismograph arrays, have indicated that most earthquakes in this region are restricted to the upper 10-15 km of the crust [Rynn and Scholz, 1978; Cowan, 1992; Reyners and Cowan, 1993].

The Hope River segment of the Hope fault (Figure 1a) and an adjacent splay (Kakapo fault) have each ruptured during historic earthquakes of magnitude $M \sim 7.3$ and $M \sim 7.0$, respectively [Eiby, 1968; Cowan, 1991; Yang, 1991, 1992]. Two more events of magnitude 6.5-6.9 have occurred in the eastern part of North Canterbury, in 1901 and 1922, and another event of M_w 6.7 occurred in June 1994, to the west of the PPAFZ. The 1994 earthquake occurred in the upper crust, and the 1901 and 1922 events are presumed to have also been shallow, based on relatively limited felt areas [McKay, 1902; Skey, 1925]. Many more faults in North

Figure 1b. Shallow seismicity (<40 km) for the region of Figure 1a during the period 1942-1994 (May 30) [New Zealand Seismological Observatory, 1994]. Symbols are scaled with magnitude, and the catalogue is complete above M 3.0 since 1988; M 4.0 since 1964; M 5.0 since 1942. The dense clusters of seismicity, north of the Hope fault and west of the PPAFZ, are associated with earthquakes of M_w 5.9 and M_w 6.7, in 1990 and 1994, respectively [Anderson *et al.*, 1993; Árnadóttir *et al.*, in press].

Canterbury exhibit evidence of Holocene displacement [Gregg, 1964], and since 1964, the New Zealand national seismograph network has located frequent earthquakes of small to moderate magnitude ($M < 5.4$) along the PPAFZ, making this one of the better defined surface fault zones in New Zealand [Reyners, 1989]. Data from the instrumental catalogue are utilized later in this paper for comparisons of recurrence parameters for large earthquakes within and between the Hope fault and PPAFZ regions.

Porter's Pass-to-Amberley Fault Zone

The PPAFZ extends east-northeast from the Southern Alps and passes into a fold and thrust belt near the town of Amberley and offshore (Figures 1a and 2) [Barnes, 1994; Nicol *et al.*, 1994]. Principal elements of the PPAFZ bound the northern margin of the Canterbury Plains where the foothills of the Southern Alps rise to 2000 m. The PPAFZ is defined by zones of crushed Torlesse greywacke that strike east and northeast, at a high angle to the regional NW-SE grain of Mesozoic deformation [Gregg, 1964], and locally cut through the late Cretaceous-Cenozoic cover sequence, disrupting late Pleistocene and Holocene landforms. A total offset of <2 km across the PPAFZ is inferred from the strike separation of

Oligocene and lower Miocene limestone beds across the Mount Grey block and vector analysis of slip across Mount Oxford (Figure 2) (H.A. Cowan *et al.*, manuscript in preparation 1995).

The Porter's Pass fault forms the principal western element of the PPAFZ and is associated with Holocene offsets near Porter's Pass [Speight, 1938; Wellman, 1953; Berryman, 1979; Coyle, 1988; Knuepfer, 1992]. New data on late Holocene faulting from this area are presented here. Farther east, the Porter's Pass fault bifurcates around Mount Oxford, and splays extend northeast and east to Lees Valley and the northern Canterbury Plains, respectively (Figure 2). A surface trace in Lees Valley is associated with scarps up to 6 m high on Holocene alluvial fans [Gregg, 1964; Garlick, 1992], and along the southern flank of the Ashley Range, major zones of crushed rock diverge to the east and southeast (Figure 2). The east striking zone extends into a region dominated by thrust faulting and folding centred on Mount Grey, whereas the southeast striking zone passes beneath the Canterbury Plains to the Cust anticline and Ashley fault and may extend offshore to the Pegasus Bay fault zone farther east [Kirkaldy and Thomas, 1963; Herzer and Bradshaw, 1985; Barnes, 1994] (Figure 1a). Data on late Holocene faulting and landslides in the eastern part of the PPAFZ have been collected from Cust anticline, Ashley fault, and Mount Grey.

Figure 2. Structural elements of the Porter's Pass-to-Amberley Fault Zone (PPAFZ) in North Canterbury, New Zealand, showing the locations of main faults, folds, and prehistoric landslides. Faults that are depicted without sense-of-movement indicators represent mapped zones of crushed rock. Numbered landslides correspond to data in Table 2 and Figure 6. Thin dashed lines denote the margins of active folds on the Canterbury Plains. Abbreviations are KA, Kowai anticline; CA, Cust anticline; MO, Mount Oxford, and MG, Mount Grey. The locations of Figures 3, 4, and 5 are indicated.

Holocene Faulting in the PPAFZ

Porter's Pass Fault

Studies near Porter's Pass, where the PPAFZ is relatively narrow, indicate a Holocene slip rate of ~ 4 mm/yr [Berryman, 1979] and 2.7–3.8 mm/yr [Knuepfer, 1992]. A few hundred meters north of Highway 73 at Porter's Pass (Figure 2), a radiocarbon date from the lower part of a peat column accumulated behind the surface trace indicates that the surface trace probably formed 2000–2500 years B.P. (Table 1).

In a highway embankment on the south side of the pass, two buried soils, each containing charcoal, are offset across the fault (Figure 3). The crosscutting relationships together with the stratigraphy indicate possibly two ruptures during the Holocene (Table 1). The older of the two soils, offset across fault 1, may reflect a rupture between 7000 and 9000 years ago. Faults 2 and 3 have displaced both soils at least once since 7000 years ago.

Cust Anticline

The Cust anticline and Ashley fault, located on a major southeast splay of the PPAFZ, are important sites for paleoseismic investigation (Figures 2, 4). The recent nature of the deformation in this region is indicated by the presence of consolidated Pliocene-early Quaternary siltstone and conglomerate which crops out to a

height of 150 m on the northern flank of Cust anticline. A steep north-to-south gradient of Bouguer gravity anomalies across the anticline has been attributed to the presence of a fault with significant throw [Kirkaldy and Thomas, 1963]. We infer that this fault is connected to the southeast striking extension of the Porter's Pass fault at Glentui River, which appears to control the flow direction of the Ashley River in the area immediately to the northwest of Cust anticline (Figure 4).

Late Quaternary uplift of Cust anticline is evident from drainage features preserved across the crest of the fold, which is capped by two late Pleistocene loess sheets [Trangmar, 1987]. Ancestral channels of the Ashley River are incised into the loess sheets (Figure 4), so the diversion of Ashley River to its present course has presumably accompanied uplift at the western margin of the anticline, possibly during the early Holocene since the youngest ancestral channel has no loess cover. The uplift was probably caused by rupture on a fault that is associated with a 1.5-km-long surface trace (Figure 4) and a large (~ 500 m) offset in basement [Kirkaldy and Thomas, 1963].

Probable late Holocene ruptures in this area are inferred from dated landslides and the timing of Ashley River downcutting along the northern flank of the anticline. Wood collected from the base of gravel deposits overlying the strath of a terrace 13 ± 12 m above the Ashley River was dated at between 2100 and 2500 calibrated radiocarbon years old (Table 1 and Figure 4), implying an average

Table 1. Evidence for Prehistoric Surface Ruptures Along the Porter's Pass-Amberley Fault Zone, North Canterbury

Fault	Latitude, Longitude	Site Description ^a	Interpretation
Porter's Pass	43.29, 171.74	Buried soils, 6900 ± 90 years B.P. and 8900 ± 110 years B.P. [Tonkin and Bashier, 1990], are offset by the Porter's Pass fault in an embankment on the south side of Highway 73 at Porter's Pass.	One or more ruptures occurred between 7000 and 9000 years B.P., and/or, one or more ruptures occurred since 7000 years B.P.
Porter's Pass	43.29, 171.74	A peat swamp is located behind the surface trace on slopes NE of Highway 73 at Porter's Pass. Charcoal in the peat at -1.30 m defines a transition from forest to grassland (N. Moar, personal communication, 1992). The interval -1.30 m to -1.40 m is dated 1720 ± 60 years B.P. (NZ 5235). The rate of peat accumulation for the interval -1.81 m to -1.30 m is 0.5 _{min} - 0.64 _{max} mm/yr. The inferred age for the base of the peat at -2.0-2.2 m is ~2000-2500 years B.P.	Formation of the surface trace predates the onset of peat accumulation. One or more ruptures since ~2500 years B.P.
Porter's Pass	43.31, 171.66 and 43.28, 171.78	Rock avalanches bury the Porter's Pass fault trace [Burrows, 1975; Coyle, 1988]. See Table 2 for ages.	Rock avalanches, possibly triggered by fault rupture, occurred between 500 and 700 years B.P.
Cust Anticline	43.26, 172.38	A strath terrace of Ashley River is 13 ± 2 m above its active floodplain on the north limb of Cust anticline. Wood from the base of overlying gravels is dated 2345 ± 58 years B.P. (NZ 7857). Calibrated radiocarbon age ^b (90% confidence interval) is 533 B.C. to 201 B.C.	The strath terrace and landslide 8b (Table 2) are of similar age. Incision of Ashley River and differential uplift of the north limb of Cust anticline has been accompanied by one or more earthquakes since ~2500 years B.P.
Ashley	43.26, 172.48	The Ashley fault trace vertically offsets a Holocene terrace of the Okuku River by 1-1.5 m. The terrace is the same height above the active Okuku River floodplain as the dated strath terrace at Cust anticline (see above). The fault trace is truncated by a younger Okuku River terrace, 4 m base to 10 m top above its present floodplain.	Rupture of the Ashley fault induced or postdated, downcutting by the Okuku River. If the average downcutting rate were similar to the Ashley River at Cust anticline, the age of the unfaulted Okuku River terrace could be ~700 years B.P.
Mount Grey	43.09, 172.4	Exposure of the Mount Grey fault in a stream is 3 m below the surface trace. Wood is dated 2330 ± 40 years B.P. (NZ 7852), and peat, impounded behind the scarp, is faulted. Younger debris flows bury the scarp. One is offset by the fault, the other is not. A tree stump in growth position on the youngest unfaulted debris flow was dated 182 ± 60 years B.P. (NZ 7853); calibrated radiocarbon age is between 1661 A.D. and 1955 A.D.	Surface rupture felled a tree beside the surface trace, between 2400 and 2100 years B.P. A later undated rupture disrupted the wood and an overlying debris flow. Soil formed on a younger unfaulted debris flow with a minimum age of 182 ± 60 years B.P.
Mount Grey	43.09, 172.54 to 43.08, 172.54	Weathering rinds on chips of Torlesse sandstone, collected along the surface trace, indicate two modal ages for site disturbance: 320 B.C. to 360 A.D. and 1550 A.D. to 1650 A.D. (69, 1991) ^c .	The modal weathering rind ages are similar to the radiocarbon dates and are consistent with evidence of rupture documented at the stream exposure.

Weathering-rind errors are calculated by summing the standard error on the calibration curve regression and fluctuations in modal rind thickness induced by measuring errors (5%); measurement errors quoted to 95% confidence were defined by remeasuring 12 late Holocene rind populations (H.A. Nicol et al., unpublished data, 1994).

^a Conventional radiocarbon age in years B.P. (A.D. 1950) using Libby half-life (5568 years). New Zealand radiocarbon dating laboratory number in parentheses.

^b Calibrated radiocarbon ages (95% confidence intervals unless otherwise stated) based on a compilation by Stuiver and Reimer [1986] of 20 year tree ring data for the period 7210 B.C. to 1950 A.D., with offset of -30 radiocarbon years as recommended by Stuiver and Pearson [1986] and Pearson and Stuiver [1986].

^c Modal age (deducted from year of observation) from thickness of weathering rinds on surface clasts of Torlesse sandstone, using method of Chinn [1981] and calibration curve of McSaveney [1992]. The sample size and year of observation are shown in parentheses.

Figure 3. Age relations between faults and buried soils exposed in a road embankment within the Porter's Pass fault zone at Porter's Pass. See Figure 2 and Table 1 for site location and description, respectively.

downcutting rate of 5.7 ± 1.4 m/kyr. The height of the terrace strath above the present floodplain presumably reflects an element of tectonic uplift and fold growth, because upstream the river is in equilibrium, whereas downstream it is aggrading [Cowan, 1992]. Whether or not the downcutting by Ashley River reflects continuous or episodic uplift is unclear, but additional data from landslides at this locality (Figure 4) suggest that the first increment of uplift was induced by coseismic rupture in this sector of the PPAFZ.

Ashley Fault

The Ashley fault offsets Pleistocene and Holocene terraces of the Okuku River east of Cust anticline and has a surface trace length of 4.5 km [Brown, 1973; Berryman, 1979] (Figure 4). The fault is downthrown to the south, with throws of 1-4 m on a late Pleistocene aggradation surface and 0.5-1.5 m on a Holocene strath terrace [Cowan, 1992]. Differences in the throw of the fault on these two surfaces suggest at least two ruptures since the formation of the late Pleistocene aggradation surface. Variations in throw which accompany slight changes in the strike of the fault, are consistent with a minor right-lateral component of displacement, but no unequivocal offsets have been documented.

The height of the Holocene strath terrace offset by the Ashley fault is similar to the dated (i.e., 13 m) strath above Ashley River, so the terraces may be of approximately equal age. The fault trace across the Okuku River strath is not significantly degraded, so the river presumably abandoned the surface prior to, or at the time of, faulting. The oldest Okuku River terrace not offset by the fault (its riser truncates the surface trace) is about 4 m above the active floodplain (Figure 4). If the average downcutting rate of the Okuku

River is similar to the Ashley River, then the unfaulted terrace formed ~600-900 years ago. The available data thus imply that the last rupture of the Ashley fault postdated 2500 years B.P. and could have accompanied the growth of Cust anticline prior to 600-900 years B.P.

Mount Grey Fault

The Mount Grey fault comprises a north dipping thrust along the south side of Mount Grey (931 m) in the eastern part of the PPAFZ (Figure 2). The fault loops around the eastern flank of Mount Grey and bifurcates, forming an oblique, left-lateral tear fault that strikes north and dips to the west, and against which the Tertiary cover sequence has been overturned [Wilson, 1963]. The fault is exposed in a number of streams and forms a prominent surface trace and hillside bench hollow, behind which scree and boulders have accumulated.

Radiocarbon and weathering-rind dates delineate two surface rupture events within the last 2500 years (Table 1). Evidence for the oldest recognized event was obtained from a stream bank exposure of the fault (Figure 2, 5), where wood and peat were preserved on the surface of a debris flow, offset across a fault scarp that is now buried 3 m below the present ground surface. The wood was not in growth position and was buried beneath younger colluvium, which in turn was offset by the fault. A soil, developed on the debris flow and containing a trace of charcoal of unknown age defined an apparent vertical offset (downthrown to the west) across the fault. We could not discern from the field relationships whether or not this buried soil was deformed or had merely developed on an exposed fault scarp at a later date. The soil was buried beneath another debris flow, also capped by a younger soil profile containing a tree stump

Figure 4. Map of the northern Canterbury Plains showing the relationships between the Cust anticline, Ashley fault, and range-bounding elements of the PPAFZ. L8a and L8b indicate locations of the radiocarbon dates from landslide 8 (Table 2 and Figure 6). Topographic contours are reproduced at 20-m and 10-m intervals (thin solid and thin dashed lines, respectively) from NZMS 270, sheets L35 and M35, by permission of the Department of Survey and Land Information, New Zealand.

in growth position. A radiocarbon date of 182 ± 60 years B.P. obtained from an outer portion of the tree stump (Table 1) indicated a minimum age for the buried soil and the last surface rupture recognized at this site.

Chips were collected from Torlesse sandstone boulders and cobbles exposed at the ground surface along the fault trace north of the stream bank locality, and weathering-rind ages for these were calculated using the method of Chinn [1981] and calibration curve of McSaveney [1992] to evaluate the timing of site disturbance. Two modal rind thicknesses were obtained and the ages (Table 1) are in good agreement with the time intervals for which surface rupture is inferred from the radiocarbon ages. The available data collectively imply surface rupture on the Mount Grey fault, between 2300–2400 years B.P. and 300–450 years B.P.

Prehistoric Landslides of the PPAFZ

Landslides provide additional data with which to assess the recurrence interval of large prehistoric earthquakes in the PPAFZ. Twenty landslides have been identified within the PPAFZ, and the ages for 10 of these have been determined from radiocarbon and weathering-rind dating (Table 2). The landslides are mostly large

($0.2\text{--}10 \times 10^6 \text{ m}^3$) rock avalanches, derived from outcrops of Torlesse basement (Figure 2). The data indicate a clustering of ages between about 500 and 700 years B.P., with others scattered back to more than 7000 years B.P. (Table 2 and Figure 6).

Landslides in the range 500–700 years B.P. are relatively uniformly distributed along the PPAFZ and are well preserved. However, there are insufficient data to confirm a similar spatial pattern of landslides at 2300 years or older, which presumably reflects the removal of landslide debris by erosion. A large landslide identified on the north limb of Cust anticline (Figure 4), which contained whole trees and an articulated skeleton of the largest species of extinct moa, *Diornis giganteus* [Cowan, 1992], was radiocarbon dated at circa 2300 years B.P. (Table 2). This age is the same as the nearby strath terrace of Ashley River (Figure 4), and we surmise that both the landslide and the onset of river downcutting accompanied coseismic uplift of Cust anticline.

To assess whether the effects of this event extended beyond the geographic limits of our study, the ages of rock avalanches from the central Southern Alps [Whitehouse and Griffiths, 1983] were recalculated using the weathering-rind calibration curve of McSaveney [1992]. Two modal ages were obtained for Southern Alps landslides (circa 1050–1400 years B.P. and circa 1700–2100 years B.P.), and these do not coincide with modal ages for landslides

Figure 5. Stratigraphy and conventional radiocarbon ages associated with samples from an exposure of the Mount Grey fault in a stream bank on the eastern flank of Mount Grey. See Figure 2 and Table 1 for site location and description, respectively.

documented in the PPAFZ (Figure 6). The ages for three landslides from the central Southern Alps do overlap with those of the PPAFZ, in the range ~2200–2450 years B.P., but fault data indicate probable rupture of the PPAFZ during that period (Table 1).

Six landslides with ages between 500 and 700 years B.P. are located within 1 km of the major faults of the PPAFZ, and two of these bury the trace of the Porter's Pass fault near Porter's Pass (Figure 2, landslides 1 and 2). By contrast, only two landslides within this age range were dated by *Whitehouse and Griffiths* [1983] in the Southern Alps farther west. Collectively, the spatial distribution of these data suggests that if the landslides were seismically triggered, they are more likely related to local rupture of the PPAFZ rather than to a regional event on the Alpine Fault, an interpretation favored by W.B. Bull (personal communication, 1994) based on recent lichenometric dating of landslides in the Southern Alps.

The interpretation of landslide distributions in paleoseismology involves two assumptions: (1) that landslides were triggered coseismically; and (2) that all landslides of a similar age were coseismic with the same event. Landslides have been triggered by all large ($M > 7$) historic earthquakes in New Zealand, but many large landslides in seismically active regions, including the Southern Alps, have not been triggered by earthquakes [e.g., *Plafker and Ericksen*, 1978; *McSaveney et al.*, 1992; *McSaveney*, 1978, 1993]. In the absence of independent evidence, the coseismic interpretation of prehistoric landslides must therefore be regarded as tentative [*Whitehouse and Griffiths*, 1983].

For both fault traces and landslides there are at least two reasons why evidence for large paleoearthquakes could be sparse. First, it must be remembered that few dating techniques can discriminate events separated by a few years or decades, a point exemplified by three historic earthquakes in northern South Island. Both the M_s 7.8 Buller and M_s 7.0 Arthur's Pass earthquakes of 1929 [*Henderson*, 1937; *Speight*, 1933; *Dowrick and Smith*, 1990] occurred within a

period of 3 months and were followed 39 years later by the M_s 7.4, M_w 7.1 Inangahua earthquake [*Adams et al.*, 1968; *Anderson et al.*, 1994]. It is doubtful whether each of these events could be differentiated from the geological record using radiocarbon dating had they occurred during prehistoric time, although dendrochronology and lichenometry could potentially discriminate the 1929 and 1968 events [e.g., *Bull et al.*, 1994]. Second, as the elapsed time since the creation of a landform increases, so too does the likelihood that it will be destroyed, modified, or buried by subsequent processes. Furthermore, periods of non deposition may conceal the occurrence of rupture events at a given site, and assumptions of continuous sedimentation are difficult to test and are rarely proven [e.g. *Doig*, 1990; *Cowan and McGlone*, 1991]. We cannot rule out periods of non deposition at any of the sites described in this study, but the temporal clustering of landslides in the range 500–700 years B.P. and their close proximity to the PPAFZ suggest that the most recent rupture of this zone probably occurred during that period. The landslides attributed to earlier events (2000–2500 years B.P. and 7000–9000 years B.P.) are much less abundant, indicating that in New Zealand's mountainous terrain, landslide data may only be useful for characterising paleoseismicity for up to 2000–3000 years B.P.

Magnitude Estimates

The dating of surface fault ruptures provides a check on the likelihood that a given group of landslides was triggered coseismically, but the estimation of magnitude is problematic in the absence of data that define the rupture dimensions, and these parameters are poorly constrained for the PPAFZ. Relationships between landslide distributions and earthquake magnitudes have been proposed based on global compilations of data from historic events [*Keefer*, 1984], but these probably do not provide an appropriate basis for estimating the magnitude of prehistoric

Table 2. Description of Dated Landslides From the Porter's Pass-Amberley Fault Zone

Locality From Figure 2	Latitude, Longitude	Landslide Volume, 10^6 m^3	Age ^a	Method ^b	Age Calendar Years ^c
1, Acheron River ^d	43.31, 171.66	6.0 two slides ^e	500 $\pm$ 69 ^e	¹⁴ C NZ 547	A.D. 1290-1515 ^f
2, Kowai River ^g	43.28, 171.78	2.0	560 $\pm$ 90	WR 60, 1988	A.D. 1340-1520
3, Kowai River ^g	43.28, 171.80	0.2	590 $\pm$ 100	WR 40, 1988	A.D. 1300-1500
4, Coal Creek	43.40, 172.00	0.02	5600 $\pm$ 62	¹⁴ C NZ 7918	4575-4335 B.C. ^f
5, Ashley Gorge	43.19, 172.14	0.3	490 $\pm$ 70	WR 75, 1991	A.D. 1430-1570
6, Lees Valley	43.11, 172.12	10.0	580 $\pm$ 90	WR 62, 1991	A.D. 1320-1500
7, Glentui River	43.21, 172.26	4.0 two slides	600 $\pm$ 100 1210 $\pm$ 190	WR 70, 1991 WR 45, 1991	A.D. 1290-1490 A.D. 590-970
8a, Cust Anticline	43.26, 172.37	0.1	7400 $\pm$ 90	¹⁴ C NZ 7854	6425-6000 B.C. ^f
8b, Cust Anticline	43.26, 172.37	3.0	2300 $\pm$ 60 2270 $\pm$ 82	¹⁴ C NZ 7855 ¹⁴ C NZ 7856	mean of two dates 200-400 B.C. ^f
9, Mount Thomas	43.15, 172.40	4.0	2467 $\pm$ 66	¹⁴ C NZ 7895	400-765 B.C. ^f
10, Mount Grey	43.09, 172.55	0.2	3430 $\pm$ 470 bimodal 4580 $\pm$ 660	WR 99, 1991	970-1400 B.C. 1930-3250 B.C.

See Table 1 footnote for explanation of weathering-rind dating errors.

^a Years before A.D. 1950, with modal weathering rind ages rounded to nearest 5 years.

^b Symbol ¹⁴C denotes conventional radiocarbon age and sample numbers; WR is weathering rinds, followed by number of rinds and year (A.D.) of measurement.

^c Symbol ¹⁴C ages (calendar years A.D./B.C.).

^d Data from Burrows [1975].

^e Data from W.B. Bull (written communication, 1994).

^f Calibrated radiocarbon ages (95% confidence levels unless otherwise stated) based on compilation by Stuiver and Reimer [1986] of 20-year tree ring data for the period 7210 B.C. to A.D. 1950, with offset of -30 radiocarbon years as recommended by Stuiver and Pearson [1986] and Pearson and Stuiver [1986].

^g Data from Coyle [1988].

earthquakes. Only the largest prehistoric rock falls may be recognized, and variations in earthquake source mechanism and regional geology could strongly influence the distribution of landslides. For example, the 1929 and 1994 Arthur's Pass earthquakes each triggered rock falls over an area comparable to many global events of similar magnitude (M. McSaveney, personal communication, 1995), but the areas affected by large landslides of the 1929 Buller and 1968 Inangahua earthquakes [Adams, 1981] are smaller than might be expected from Keefer's global catalogue. The Buller and Inangahua earthquakes occurred in a region composed largely of crystalline Paleozoic rocks, in contrast to the weakly metamorphosed greywacke of the Southern Alps. These differences, together with variations in focal depth, rupture directivity, and topographic effects illustrate some of the uncertainties that would accompany the estimation of earthquake magnitudes from prehistoric landslide distributions. The uncertainties naturally would be compounded by incomplete sampling of the spatial distribution of prehistoric landslides.

The threshold magnitude for surface rupture and coseismic landsliding in the PPAFZ is unclear. However, since all historic earthquakes larger than magnitude $M \sim 7$ in northern South Island have been accompanied by landslides [McKay, 1902; Henderson, 1937; Speight, 1933; Adams *et al.*, 1968] we infer that the late Holocene landslides in the PPAFZ were probably triggered by earthquakes of similar magnitude. More precise estimates of magnitude would depend critically on detailed knowledge of the sub-surface fault dimensions and segmentation [Wells and Coppersmith, 1994], parameters that are poorly defined in this hybrid strike-slip and thrust fault zone. Indeed, lateral variations in fault behavior are to be expected, and historic earthquakes elsewhere [e.g., Amaiike, 1987; Stein and Yeats, 1989; Hull, 1990; Stein and Ekström, 1992; Smalley *et al.*, 1993; Anderson *et al.*, 1994; Jibson *et al.*, 1994] suggest that not all Holocene ruptures in the PPAFZ would necessarily produce surface fault offsets indicative of earthquake magnitude or amenable to paleoseismic investigation. This seems especially likely in the fold and thrust belt

Figure 6. Temporal distribution of landslides and fault ruptures in the PPAFZ, and a comparison with dated landslides from the Southern Alps [Whitehouse and Griffiths, 1983]. Weathering-rind ages from the Southern Alps have been recalibrated using the calibration curve of McSaveney [1992]. Landslide numbers correspond to those in Table 2 and are located on Figure 2. Abbreviations are A, Ashley fault; CA, Cust anticline; MG, Mount Grey fault; PP, Porter's Pass fault. Asterisks denote fault ruptures dated by weathering rinds. Refer to Tables 1 and 2 for details of fault ruptures and landslides, respectively.

at the eastern end of the PPAFZ and farther north, where late Holocene surface traces are preserved, but locally display vertical offsets that are anomalously large with respect to fault length and for which no landslides have been correlated [Cowan, 1994; A. Nicol et al., unpublished data, 1991]. Uncertainties in the assessment of fault dimensions and the likelihood of hidden faults in this young fault zone also complicate the estimation of slip rate and potential slip per event. Local variations in fault kinematics are evident from the differential uplift at Cust anticline (this study) and in the fold and thrust belt farther east [Nicol et al., 1994]. Davis and Namson [1994] and Shaw and Suppe [1994] have shown that detailed structural analysis can reveal the presence of hidden faults, and similar work in progress in the PPAFZ (H.A. Cowan et al., manuscript in preparation, 1995) may provide further insight into the kinematics and subsurface structure of the zone, especially if combined with recent observations of geodetic strain (C.F. Pearson et al., Strain distribution across the Australian-Pacific plate boundary in the central South Island, New Zealand, from 1992 GPS and earlier terrestrial observations, submitted to *Journal of Geophysical Research*; hereinafter Pearson et al.; submitted manuscript, 1995). Here, we restrict our discussion to the implications of available historical and paleoseismic data for analysis of seismicity rates and fault behavior.

Seismicity Rate Analysis

The traditional approach to seismicity rate analysis is to assume a uniform spatial distribution of earthquakes and an exponential distribution of magnitudes of the Gutenberg-Richter form:

$$\log N(M) = a - bM \quad (1)$$

where $N(M)$ is the cumulative number of earthquakes of magnitude equal to or greater than (M); a defines the rate of occurrence, and b is the exponential decay with increasing magnitude, implying self-similarity [e.g., King, 1983]. The parameters a and b are constants of the seismicity model and are computed from a catalogue of historical seismicity. Equation (1) may be integrated to obtain a cumulative exponential distribution, truncated at a maximum magnitude for a given set of faults or volume of crust, such that

$$N = N_0 [10^{b(M_0 - M)} - 10^{b(M_0 - M_{\max})}] \quad (2)$$

N_0 is the number of earthquakes of magnitude M_0 or greater, where M_0 is an arbitrary threshold magnitude [e.g., Youngs and Coppersmith, 1985]. When combined with an appropriate attenuation expression, the seismicity model may be used to derive return periods or probabilities of exceedance for specified levels of ground motion. This classical hazard analysis, advanced by Cornell [1968], has been applied widely in seismic hazard assessment [see J.G. Anderson et al., 1993] and in New Zealand by, among others, Peek et al. [1980] and Smith and Berryman [1986].

Studies of individual faults have shown that the recurrence rate of large earthquakes ($M > 7$) may be under-predicted by the b value model [e.g., Wesnousky et al., 1983; Schwartz and Coppersmith, 1984; Davison and Scholz, 1985]. Rather, the frequency of these large events appears to be governed by relations between fault geometry, segmentation, and cumulative geological offset, which influence the distribution of strength properties along fault zones

[e.g., King and Nabelek, 1985; Sibson, 1989; Wesnousky, 1988, 1990]. The recognition of structural controls on the propagation, arrest, and subsequent slip distribution associated with some coseismic fault ruptures [e.g., Sibson, 1985, 1989] has drawn attention to the significance of fault segmentation as a possible indicator of the size of earthquakes associated with a given fault. This concept is implicit in the model of "characteristic" earthquake recurrence in which most of the deformation associated with a given fault or segment is released coseismically by earthquakes that occur within a relatively narrow recurrence time and magnitude distribution [Working Group on California Earthquake Probabilities, 1990]. Other models advanced to explain geological evidence of time-variable and slip-variable fault behavior [e.g., Schwartz, 1989; Scholz, 1989] also imply the nonrandom occurrence of large earthquakes.

There is considerable debate as to how widely and over what timescale the "characteristic earthquake" and other models of periodic fault behavior may apply [cf. Nishenko, 1991; Nishenko and Sykes, 1993; Kagan and Jackson, 1994, 1995; Roeloffs and Langbein, 1994]. While the behavior of some faults appears to be governed by characteristic earthquakes over several earthquake cycles [e.g., Schwartz and Coppersmith, 1984] and others by temporal clusters of earthquakes [e.g., Wallace, 1987; Coppersmith, 1988; Sieh et al., 1989], comparisons of seismicity at the scale of plate boundary segments have indicated nonperiodic fault behavior [Kagan and Jackson, 1995].

Recent analysis of historic seismicity and geological fault-slip data from southern California [Wesnousky, 1994] indicates that the structurally mature faults of the San Andreas fault zone do exhibit characteristic behavior, at least over the time periods (10^2 – 10^4 years) considered. In North Canterbury, one historic surface rupture of the Hope River segment of the Hope fault (Figure 1a), together with paleoseismic data and spatial variations in late Quaternary slip rate from the same segment, are also consistent with the characteristic earthquake model [Cowan, 1990, 1991; Cowan and McGlone, 1991]. Elsewhere in New Zealand, significant variations in fault behavior have been documented [Berryman and Beanland, 1991], and the problem there, as in most parts of the world, is one of reconciling geological and seismological data spanning different time periods. The seismicity catalogue for New Zealand spans only the last 150 years for large earthquakes, and the instrumental catalogue spans a much shorter period. This problem is intractable and acknowledged in seismicity rate analysis and hazard estimation. However, the historic catalogue for North Canterbury is sufficiently long to illustrate significant discrepancies between recurrence rates for large earthquakes extrapolated from the historic catalogue versus recurrence rates implied by geological data.

Frequency-Magnitude Relations for North Canterbury Seismicity

Seismicity Zones

For the purpose of comparing historic seismicity and paleoseismic data, two 100×50 km zones (Figures 1a and 1b) were selected, based on the contrasting structural maturity and geological slip rates of the known faults. Zone 1 encompasses an ~35-km-long releasing bend in the Hope fault zone across which about half (~10–14 mm/yr) of the total Hope fault slip rate is localized on the Hope River segment, with the remainder distributed across adjacent plays, notably the Kakapo fault [Cowan, 1990; Van Disen and Yeats, 1991; Yang, 1991] (Figure 1a). The Hope fault has a cumulative offset of ~20 km [Freund, 1971], and its total slip rate

(~25 mm/yr) is equivalent to more than 60% of the total rate of relative plate motion [DeMets et al., 1990], which represents the highest rate of slip in the Marlborough Fault System [Van Disen and Yeats, 1991]. The last surface rupture in zone 1 occurred on the Hope River segment during an $M \sim 7.3$ earthquake in 1888 (Figure 1a) [McKay, 1890; Cowan, 1991]. The Hope River segment has ruptured repeatedly at intervals of 80–200 years during the last millenium, or approximately every 140 years during the last 3500 years, if the slip rate and local offset in the 1888 event are included [Cowan and McGlone, 1991] (Table 3).

Zone 2 is centered on the PPAFZ whose cumulative offset and slip rate is approximately one order of magnitude smaller than the Hope fault. Both zone 1 and zone 2 contain additional faults associated with Holocene surface traces, for which little or no paleoseismic data are available [Wood et al., 1994; this study]. Uncertainties in the location of historic earthquakes smaller than magnitude $M 4.0$, preclude the selection of smaller zones adjacent to the major faults [cf. Wesnousky, 1994]. However, the depth distribution of seismicity recorded during microearthquake surveys in zone 1 [Arabasz and Robinson, 1976; Kieckhefer, 1977] and zone 2 [Reyners and Cowan, 1993] indicates that most seismicity is restricted to the upper 10–15 km of the crust. By conservatively choosing 50 km wide zones for seismicity rate analysis in this study, we are therefore confident that all seismicity associated with the major seismogenic faults has been sampled.

Seismicity Distribution and Recurrence Parameters

Three time periods were selected based on the relative completeness of the seismicity catalogue: for events $M > 5$ since 1942; $M > 4$ since 1964 and $M > 3$ since 1888. Only those events with crustal depths (i.e., restricted and free depths shallower than 4 km) were sampled, so as to exclude events located in the subducted Pacific plate 100 km beneath the Hope fault region [Robinson, 1991]. The catalogue was also truncated at May 30, 1994, to avoid sampling those aftershocks of the June 1994 Arthur's Pass earthquake that fall within the northwestern corner of zone 2 (Figure 1b).

The instrumental seismicity catalogue for zone 1 contains a number of events of magnitude $M 4.0$ – 6.4 , mainly in the east. Several of these events are associated with two swarms, the first of which occurred south of the Hope fault on May 22, 1948, and comprised six events ($M_L = 5.9, 6.4, 6.2, 5.7, 5.7$, and 5.8) within the space of 2 hours and 24 min [Eiby, 1982]. The second swarm (April 22–28, 1973) consisted of four events ($M_L = 5.2, 4.5, 4.4$, and 5.0) in an area northeast of the Hope fault [Reyners, 1989]. Only the largest event of that swarm lies within the northeastern corner of zone 1. Neither of the two swarms can be attributed to movement on known faults, but the Seismological Observatory File locations suggest that they are probably unrelated to the Hope fault.

One interesting feature of the catalogue for zone 1 is the absence of events of any magnitude in the western part of the region and farther west, where the 1888 and 1929 ruptures occurred (Figure 1b). This quiescence was also apparent during an earlier study of microseismicity [Arabasz and Robinson, 1976] and may reflect a significant reduction of stress in that region following the large historic earthquakes.

The seismicity catalogue for zone 2 (Figure 1b) clearly shows diminishing activity south of the PPAFZ, but the strike of that zone is remarkably well-defined as noted by Reyners [1989], despite all but one event ($M \sim 6.2$) [Eiby, 1990], being smaller than $M 5.5$. A recent study of microseismicity in North Canterbury [Cowan, 1992; Reyners and Cowan, 1993; H.A. Cowan et al., manuscript in preparation, 1995] has also shown seismicity clustered along the

Table 3. Magnitude-Frequency Relations Within the Hope Fault Region (Zone 1) and Porter's Pass-to-Amberley Fault Zone (Zone 2)

Period (Number of Years)	Threshold Magnitude	Number of Events Regressed	Recurrence Parameters				Return Periods (years)		
			<i>a</i>	SE	<i>b</i>	SE	<i>M</i> 6 ^S	<i>M</i> 7 ^S	<i>M</i> > 7 ^P
<i>Hope Fault Region (Zone 1)</i>									
1942-1994	M 5	10	3.76	0.47	0.87 ^a	0.08	29	214	80-200 ^b
(52.5)			3.64	0.47	0.85	0.08	29	204	
1964-1994	M 4	4	3.31	0.22	0.87 ^a	0.05	81	602	60-300 ^c
(30.5)			3.22	0.22	0.85	0.05	76	537	
1988-1994	M 3	32	4.33	0.23	1.20 ^a	0.06	-	-	
(6.5)			3.14	0.43	0.87	0.12	120	891	
<i>PPAFZ Region (Zone 2)</i>									
1942-1994	M 5	6	1.83	0.75	0.60	0.13	59	234	1300-
(52.5)			3.33	0.88	0.87	0.16	78	575	2000 ^d
1964-1994	M 4	20	3.01	0.35	0.81	0.07	71	457	750-1000 ^e
(30.5)			3.29	0.36	0.87	0.08	85	631	
1988-1994	M 3	48	3.81	0.13	0.97 ^a	0.04	102	955	1500-2000 ^f
(6.5)			3.43	0.17	0.87	0.05	62	457	

Except as otherwise noted, *b* was determined through an iterative procedure in which only *a* was allowed to vary. Two sets of recurrence parameters are given for each of the three time periods in order to evaluate the sensitivity of *a* and *b* values. The standard errors (SE) indicate no significant differences in *a* for different values of *b*. The superscript S and P denote historical return periods implied by seismicity and paleoseismic data, respectively.

^a The *a* and *b* cases were determined simultaneously by a linear least squares fit.

^b Cowan and McGlone [1991].

^c Calculated using slip rate of 10-25 m/kyr and slip per event of 1.5-3.0 m. Data are from Cowan and McGlone [1991] and Van Dissen and Yeats [1991].

^d From landslide and fault rupture data (this study).

^e Calculated using a slip rate of 3-4 m/kyr and slip per event of 3 m. Slip rates from Berryman [1979] and Knuepfer [1992].

^f Calculated using a slip rate of 3-4 m/kyr and slip per event of 6 m. Slip rates from Berryman [1979] and Knuepfer [1992].

PPAFZ, with most events confined to the upper 8-15 km of the crust.

The recurrence parameters *a* and *b* of the Gutenberg-Richter relation were computed for the PPAFZ and Hope fault regions using ordinary least squares regression and an iterative approach to the selection of *b* values (Table 3). The recurrence parameters listed in Table 3, and illustrated for the period 1964-1994 in Figure 7, indicate negligible differences in activity rates between the selected regions, despite an order of magnitude difference in slip rate and cumulative offset. By contrast, the paleoseismic data from each region imply significant differences in recurrence intervals for large earthquakes (Table 3), yet neither could be reliably predicted from the historic seismicity catalogue.

In zone 1 the presence of many additional faults with potential to generate large earthquakes in the Hope fault region emphasizes the deceptiveness of contemporary seismicity. Moreover, the discrepancy between recurrence rates for large earthquakes inferred from historical and geological data would be accentuated if the swarm earthquakes mentioned above were to be excluded from the analysis; i.e., the seismicity rate above magnitude 5 would be even lower in zone 1 during the historical period. The situation is different in zone 2, where the discrepancy between historic and paleoseismic data is in the opposite sense (Table 3). There, the recurrence rate for large earthquakes is approximated by the

Gutenberg-Richter model (Figure 7), but the closest match between the historical and geological data and the smallest errors for the historical recurrence parameters are obtained from the magnitude-frequency distribution of *M* > 3 earthquakes since 1988 (Table 3). The recurrence estimates for large earthquakes derived from longer periods of historical seismicity in zone 2 are higher than those implied by the paleoseismic data (Table 3).

Discussion

Further interpretation of the historical seismicity data may be unwarranted given that the seismicity catalogue is possibly too short to sustain a robust "productivity" analysis at magnitudes less than *M* 4. The historical data in the magnitude range *M* < 6.0 straddle a period of seismic quiescence in the Hope fault region, which may not be representative of the longer-term seismicity rate. Similarly, the relatively high rate of seismicity in the PPAFZ during the last 30 years may reflect long-term temporal variations in deformation and clustering of earthquake occurrence, as for example, inferred from comparisons of low geological slip rate but higher geodetic strain and seismicity rates, in the reverse fault province of northwestern South Island [H. Anderson et al., 1993; Beavan et al., 1994].

Figure 7. Comparison of frequency-magnitude relations for the Hope fault (Zone 1) and PPAFZ (Zone 2) (Figure 1b) for the period 1964-1994. The frequency histograms were compiled at class intervals of M 0.1. The solid line represents the cumulative frequency, whereas the dashed straight line denotes the best fit (b value 0.87) for the magnitude range defined by vertical ticks. The shaded boxes indicate the recurrence rate and possible magnitude for large earthquakes inferred from geological data. The magnitude-frequency distribution for the Hope fault is in accord with the characteristic earthquake model, whereas that of the PPAFZ is approximated by the Gutenberg-Richter model. Note in Table 3, however, that the Gutenberg-Richter relationship overpredicts the recurrence rate of $M > 7$ earthquakes in the PPAFZ.

Temporal variations in seismicity rate may profoundly affect the frequency-magnitude distribution and apparent relationships between seismic productivity and fault slip rate, structural complexity, or maturity. For example, had it been possible to sample seismicity in the Hope fault region for the 50 years straddling the 1888 earthquake, the productivity could have been higher and the frequency-magnitude relationships implied by such data might not have underpredicted the recurrence rate for the largest events on the Hope fault to the same degree as the current data set (Table 3). Conversely, low seismicity rates may typify the behavior of the Hope fault, since it is clear that the $M \sim 7.3$ event of one century ago has not been followed by ongoing high rates of activity.

Return periods for large rare events are clearly sensitive to slight changes in recurrence parameters when regressions are performed on data dominated by smaller events, or when the catalogue is heterogeneous with respect to magnitude [e.g., Okal and Romanowicz, 1994; Kagan and Jackson, 1995]. It is perhaps ironic therefore that the closest match between the historical and geological data from the PPAFZ should be obtained from only 6.5 years (1988-1994) of seismicity data (Table 3), given that recurrence estimates based on such a short time period would be assigned a low weighting in any seismic hazard assessment of that region.

The available paleoseismic data show that even at this early stage of development, elements of the PPAFZ are amenable to paleoseismic investigation. The apparent absence of large landslides or surface ruptures with ages between or younger than 500-700 and 2000-2500 years B.P. indicates that the data for the late Holocene may be complete. However, it is clear from the distribution of landslide ages that evidence for events older than

~ 2500 years is only sparsely preserved. Moreover, while the combined evidence of faulting and landslides between 2000 and 2500 years B.P. is quite compelling, this cannot be said of the 500-700 year B.P. event, for which our interpretation rests largely on the landslide data alone.

If our record of late Holocene rupture of the PPAFZ is complete, the implied interval of 1300-2000 years between the last two events would require 4-8 m displacement per event to be consistent with the inferred slip rate of 3-4 mm/yr on the Porter's Pass fault. Displacements per event of this order have been documented for numerous Holocene faults throughout New Zealand [Berryman and Beanland, 1991] and Árnadóttir et al. [1995] have modeled up to 5 m of displacement for the 1994 Arthur's Pass earthquake (Figure 1a). Thus the inferred return period, displacement per event, and slip rate need not be inconsistent. Moreover, it seems reasonable to conjecture that larger displacements per event could be expected within a newly formed fault zone composed of numerous short (<10 km) segments that might accommodate more elastic strain than a mature fault which has undergone strain softening. The behavior of the PPAFZ at its present stage of development could therefore be more consistent with that of intra-plate fault zones, i.e., characterized by low slip rate, more interseismic fault healing processes, and larger slip or stress drop per event [Kanamori and Allen, 1986].

Although the available recurrence and displacement estimates for large earthquakes in the PPAFZ are consistent with the inferred geological slip rate of the Porter's Pass fault, we cannot discount the possibility that the slip per event is less than that required to account for the total strain, which recent studies have shown to be geodetically measurable at well over the 95% confidence level

(Pearson et al., submitted manuscript, 1995). At present, slip rates for faults in the central and eastern parts of the PPAFZ are too poorly defined to permit moment rate summation for direct comparison with rates of geodetic strain, but it is possible that some fraction of the slip across the PPAFZ is distributed among secondary structures, or accommodated by smaller earthquakes than we presently recognize in the geological record. One of several explanations advanced to explain a historic deficit of moment release in California [e.g., *Dolan et al.*, 1995] is that a larger fraction of the total slip rate across that region is released by more frequent moderate ($M \sim 6-6.5$) earthquakes for which there is less geological evidence [Ward, 1994; Wesnousky, 1994]. Alternatively, a proportion of the strain may be accommodated by relatively aseismic slip in numerous small earthquakes [e.g., *Lin and Stein*, 1989; *Hill et al.*, 1990]. In the PPAFZ, the historical seismicity rates may be "randomly" high for the sampled time periods and magnitude range considered, but it is also plausible that the frequency of large earthquakes may be quasi-periodic and governed by factors unique to a fault zone in the early stages of development.

Conclusions

Dated surface ruptures and landslides corroborate an inference of two large ($M > 7.0$) earthquakes within the PPAFZ during the last ~2500 years. Two earlier events during the Holocene are also recognized, but the data prior to 2500 years are presumed to be incomplete. A return period of 1300-2000 years for rupture of the PPAFZ would be consistent with the inferred slip rate (3-4 mm/yr) of its principal element, the Porter's Pass fault, provided that the slip per event is in the range 4-8 m.

Paleoseismic data from the adjacent structurally mature Hope fault zone farther north indicate significant differences in fault behavior with respect to the PPAFZ, with recurrence intervals for large earthquakes apparently differing by a factor of 5 or more. Paradoxically, the differences in historical seismicity rates for the Hope fault and PPAFZ (zones 1 and 2, respectively, Figure 1b) are negligible despite an order of magnitude difference in slip rate and cumulative offset associated with the major faults. The magnitude-frequency distribution for the Hope fault region is in accord with the characteristic earthquake model, whereas the magnitude-frequency distribution in the PPAFZ is approximated, but slightly overpredicted, by the Gutenberg-Richter relationship (Table 3 and Figure 7). The comparison of these two fault zones demonstrates the importance of the structural maturity of the fault zone in relation to seismicity rates inferred from recent, historical, and paleoseismic data.

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