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Final Report to the Earthquake Commission on Project No. 6UNI 502: 'A Geodetic Investigation of Slow Slip in the Hikurangi Subduction Zone Beneath Raukumara Peninsula, New Zealand'

Anne Douglas, John Townend, John Beavan, Laura Wallace.

September 27, 2005

1 Layman's abstract

Fault slip episodes, occurring over much longer time periods (days–months) than earthquakes, have been recorded using Global Positioning System (GPS) instruments at subduction margins around the Pacific Rim. These so-called slow slip events may make a significant contribution to moment release in subduction zones, so estimating their size and frequency is a key task in characterizing seismic hazard at subduction zones.

In October 2002, a slow slip event of 20–30 mm surface displacement was observed on two continuous GPS instruments near Gisborne on the Raukumara Peninsula, North Island, New Zealand. This event was the first of its kind to have been documented with continuous GPS in New Zealand. Scientists from the Institute of Geological and Nuclear Sciences (GNS) interpreted the motion to have been caused by slip along the boundary between the Pacific and Australian tectonic plates under the seafloor beneath Poverty Bay.

As continuous GPS coverage of the Gisborne region was sparse in 2002, it has not been possible to quantify the extent of the October 2002 slow slip event using continuous GPS data alone. In this project, we have used regional campaign GPS data (collected during day-long deployments at 2–3 yr intervals over the past ten years) in combination with the available continuous GPS data to study the October 2002 slow slip event and to find out whether other such events may have occurred in the past ten years.

Due to infrequent sampling in the campaign GPS data we cannot constrain the spatial extent and amount of slip during the October 2002 event any further than with the continuous GPS data. However, we have successfully used campaign and continuous GPS time series from neighbouring stations to make recurrence interval estimates for events of similar surface displacement to the 2002 event in the Gisborne region. Our calculations show that such events recur at 2–4 year intervals, which is in agreement with the occurrence of the November 2004 slow slip event in the same region as the 2002 one. Our studies using regional data sets also indicate that the effects of the Gisborne 2002 slow slip event were localised.

We have made a preliminary study into the possible association of seismic tremor with the slow slip events recorded on the Raukumara Peninsula, such as has been found in North America and Japan. We find evidence for tremor recorded on multiple stations across the Raukumara Peninsula broadband seismic network during both the 2004 Gisborne slow slip event, but we have not rigorously examined the spatial and temporal relationship of this tremor to the slow slip event.[1]

2 Technical abstract

In October 2002, a slow slip event of 20–30 mm surface displacement was observed on two continuous Global Positioning (GPS) instruments near Gisborne on the Raukumara Peninsula, North Island, New Zealand. This event was the first of its kind to be documented with continuous GPS in New Zealand.

We have processed and analysed nearly ten years of regional campaign GPS records (1995–2004), which we use, in conjunction with recent continuous GPS and broadband seismic data, to study the spatial extent of the Gisborne 2002 slow slip event and its effect on regional deformation. We find that infrequent sampling in the campaign GPS data set has aliased the Gisborne 2002 slow slip event and we cannot constrain the spatial extent and moment release of this event any further than with the continuous GPS data. We successfully use campaign and continuous GPS time series from neighbouring stations to make estimates of the recurrence interval for events of similar surface displacement to the 2002 slow slip occurring

[1]Seismic tremor will be the focus of a Marsden-funded project led by John Townend, commencing in early 2006.

in the Gisborne region. Our calculations reveal that such events recur at 2–4 year intervals, which is in agreement with the recent November 2004 slow slip event in the Gisborne region. Campaign GPS time series and regional deformation fields show evidence for more than one slow slip event occurring in the Gisborne area prior to 2001. Forward modelling shows that the surface displacements recorded during the Gisborne 2002 slow slip event are the result of 18 cm of slip on a plane, ca. 60 km $\times$ 25 km, on the subduction interface offshore of Gisborne, though the fault length is poorly constrained. The Gisborne 2002 slow slip event is shown to have had no significant effect on inter-seismic coupling and slip deficit on the Hikurangi subduction margin, or on regional deformation patterns over the Raukumara Peninsula. Our model of slip suggests that slow slip beneath the Raukumara Peninsula currently occurs in episodic events involving local slip at the base of the seismogenic zone on the subduction interface offshore of Gisborne. Our studies using regional data sets also indicate that effects of the Gisborne 2002 slow slip event were localised.

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[2]Seismic tremor will be the focus of a Marsden-funded project led by John Townend, commencing in early 2006.

3 Publications/outputs relating to this project

1. Douglas, A., Beavan, J., Wallace, L., Townend, J., 2005. Slow slip on the northern Hikurangi subduction interface, New Zealand. Geophys. Res. Lett., Vol. 32, No. 16, L16305, doi:10.1029/2005GL023607.

2. Douglas, A., 2005. A geodetic investigation of slow slip in the Hikurangi subduction zone beneath Raukumara Peninsula, New Zealand. Unpublished M.Sc. thesis, Victoria University of Wellington, 118 pp.

3. Beavan, J., Wallace, L., Fletcher, H., Douglas, A., Townend, J., 2005. Aseismic slip on the Hikurangi subduction interface, New Zealand. IAG/-IAPSO/IABO Joint Assembly August 2005, Abstract.

4. Douglas, A., Beavan, J., Wallace, L., Townend, J., Smith, E., 2004. Aseismic slip on the Hikurangi subduction zone, New Zealand. Eos Trans. AGU, 85(47), Fall Meet. Suppl., Abstract S42B-03.

5. Douglas, A., Beavan, J., Wallace, L., Townend, J., Smith, E., 2004. Investigating Aseismic Slip Beneath the Raukumara Peninsula, North Island, New Zealand. In: Manville, V., and Tilyard, D. (eds.), Programme and Abstracts, Geological Society of New Zealand/New Zealand Geophysical Society/26th Annual Geothermal Workshop combined conference "GEO3", Taupo, New Zealand. Geological Society of New Zealand Misc. Publ. 117A.

A Geodetic Investigation of Slow Slip in the Hikurangi Subduction Zone Beneath Raukumara Peninsula, New Zealand

Anne Douglas

31 March 2005

A thesis submitted to the School of Earth Sciences,
Victoria University of Wellington,
for the degree of Master of Science in Geophysics.

Abstract

In October 2002, a slow slip event of 20–30 mm surface displacement was observed on two continuous Global Positioning (GPS) instruments near Gisborne, North Island, New Zealand. This event was the first of its kind to be documented with continuous GPS in New Zealand.

We have processed and analysed nearly ten years of regional campaign GPS records (1995–2004), which we use, in conjunction with recent continuous GPS and broadband seismic data, to study the spatial extent of the Gisborne 2002 slow slip event and its effect on regional deformation. We find that infrequent sampling in the campaign GPS data set has aliased the Gisborne 2002 slow slip event and we cannot constrain the spatial extent and moment release of this event any further than with the continuous GPS data. We successfully use campaign and continuous GPS time series from neighbouring stations to make estimates of the recurrence interval for events of similar surface displacement to the 2002 slow slip occurring in the Gisborne region. Our calculations show that such events recur at 2–4 year intervals, which is in agreement with the recent November 2004 slow slip event in the Gisborne region. Campaign GPS time series and regional deformation fields show evidence for more than one slow slip event occurring in the Gisborne area prior to 2001. Forward modelling shows that the surface displacements recorded during the Gisborne 2002 slow slip event are the result of 18 cm of slip on a plane, ca. 60 km $\times$ 25 km, on the subduction interface offshore of Gisborne, though the fault length is poorly constrained. The Gisborne 2002 slow slip event is shown to have had no significant effect on inter-seismic coupling and slip deficit on the Hikurangi subduction margin, or on regional deformation patterns over the Raukumara Peninsula. Our model of slip suggests that slow slip beneath the Raukumara Peninsula currently occurs in episodic events involving local slip at the base of the seismogenic zone on the subduction interface offshore of Gisborne. Our studies using regional data sets also indicate that effects of the Gisborne 2002 slow slip event were localised.

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The project was fully funded by the Earthquake Commission (Project Number 6UNI502). Geological and Nuclear Sciences (GNS) in Lower Hutt provided me with an office space and access to archived data. Fieldwork for the 2004 GPS campaign was organised and lead by Neville Palmer, who made sure that we had a great new data set to work with.

Thank you to the GNS staff in MacKay wing, who were always welcoming and interested in my progress. In particular, my project benefitted from discussions with Martin Reyners and Kelvin Berryman. Kevin Fenaughty and Mark Chadwick, from GeoNet, facilitated convenient access to broadband seismic data, and Mike Hagerty helped to process and interpret that data. Thanks also to the staff and students in the Geophysics department. Ralph Wahrlich has done a great job at helping me with computer problems throughout the year. I enjoyed the company of my officemates Natalie Balfour, Nick Horspool, Vasso Mouslopoulou, Thessa Tormann and Jo Whittacker, and appreciated the help and inspiration they gave me throughout the year.

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To my friends and family in Wellington, and in the South, thanks for keeping me sane. Thanks to James for helping with maths (and statistics) questions, and the LaTeXing. Thanks to Dad for proofing the bibliography; and thanks to Mum, Dad, Helen and James for asking questions and being interested.

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Contents

List of Figures

List of Tables

Chapter 1

Introduction

Over the past fifteen years, dense networks of continuously recording global positioning system (GPS) sites around the world have greatly illuminated spatial and temporal patterns in earth deformation. Short-term earth deformation processes occurring over time periods of days to years can now be studied in detail using highly accurate GPS technology. In particular, this has lead to the recognition of a form of tectonic strain release in which episodes of fault slip occur over weeks or months. These 'slow slip events', which resemble earthquakes but occur over much longer time-frames, have been recorded in subduction zones around the Pacific rim, including the Hikurangi margin in the northeastern North Island of New Zealand. As subduction zones generate the world's largest earthquakes, understanding the process of slow slip and its contribution to local strain accumulation, or release, is vital in the accurate assessment of seismic hazard in these actively deforming regions.

1.1 Statement of objectives

The aim of this project is to identify and characterise the deformation associated with slow slip events beneath the Raukumara Peninsula on the northern Hikurangi subduction margin. This will involve four main tasks:

1. Re-analysis of existing GPS data collected during campaigns in 1995, 1997, and 2001, using modern software and uniform satellite orbits;

2. Collection and analysis of new campaign GPS data from a subset of Raukumara Peninsula sites;

3. Interpretation of the combined 1995–1997–2001–2004 geodetic data set, allowing for the fact that there may be slow slip events represented in the data, and with specific knowledge of slow slip events which have been recorded since 2002;

4. Preliminary analysis of continuous broadband seismic data to investigate the possible association of seismic tremor with slow slip events recorded geodetically on the Raukumara Peninsula.

In this chapter we establish a regional and global context for the slow slip events recorded since 2002 with continuous GPS on the northern Hikurangi subduction margin. We review relevant work done previously in the Raukumara region, an important location for land-based observations of the shallow subduction margin. Studies in geology, neotectonics, geodesy and seismicity have all been undertaken in this area; this study builds on previous work as we seek to understand the phenomenon of slow slip and deformation along the northern Hikurangi subduction margin. We also examine three case studies of slow slip deformation on the subduction interface from around the Pacific rim. The Gisborne 2002 slow slip event is one of many such events to have been recorded from subduction zones around the Pacific over the past ten years. The case studies we present are well developed in the literature and provide a broader tectonic context for the slow slip events observed beneath the Raukumara Peninsula.

1.2 Slow slip beneath the Raukumara Peninsula

In October 2002, a slow slip event was recorded on two continuous GPS stations in the Gisborne area (Figure 1.1, Beavan et al., 2003). This event was the first of its kind to be recorded by continuous GPS in New Zealand, and is the starting point for this thesis. During the 12-day duration of the slow slip event, surface displacements of approximately 20 mm east, 5 mm south and 5 mm downwards were recorded at station GISB in the GeoNet[1] continuous GPS network. GIS1, a privately run GPS station closer to the coast, recorded a similar signal with an eastward displacement of about 30 mm. Beavan et al. (2003) inferred the cause of this event to be reverse slip of about 20 cm on the Hikurangi subduction interface offshore from Poverty Bay. The known occurrence of slow slip deformation beneath the Raukumara Peninsula offers a new opportunity to re-examine the ten year span of campaign GPS data from the region

[1] http://www.geonet.org.nz/

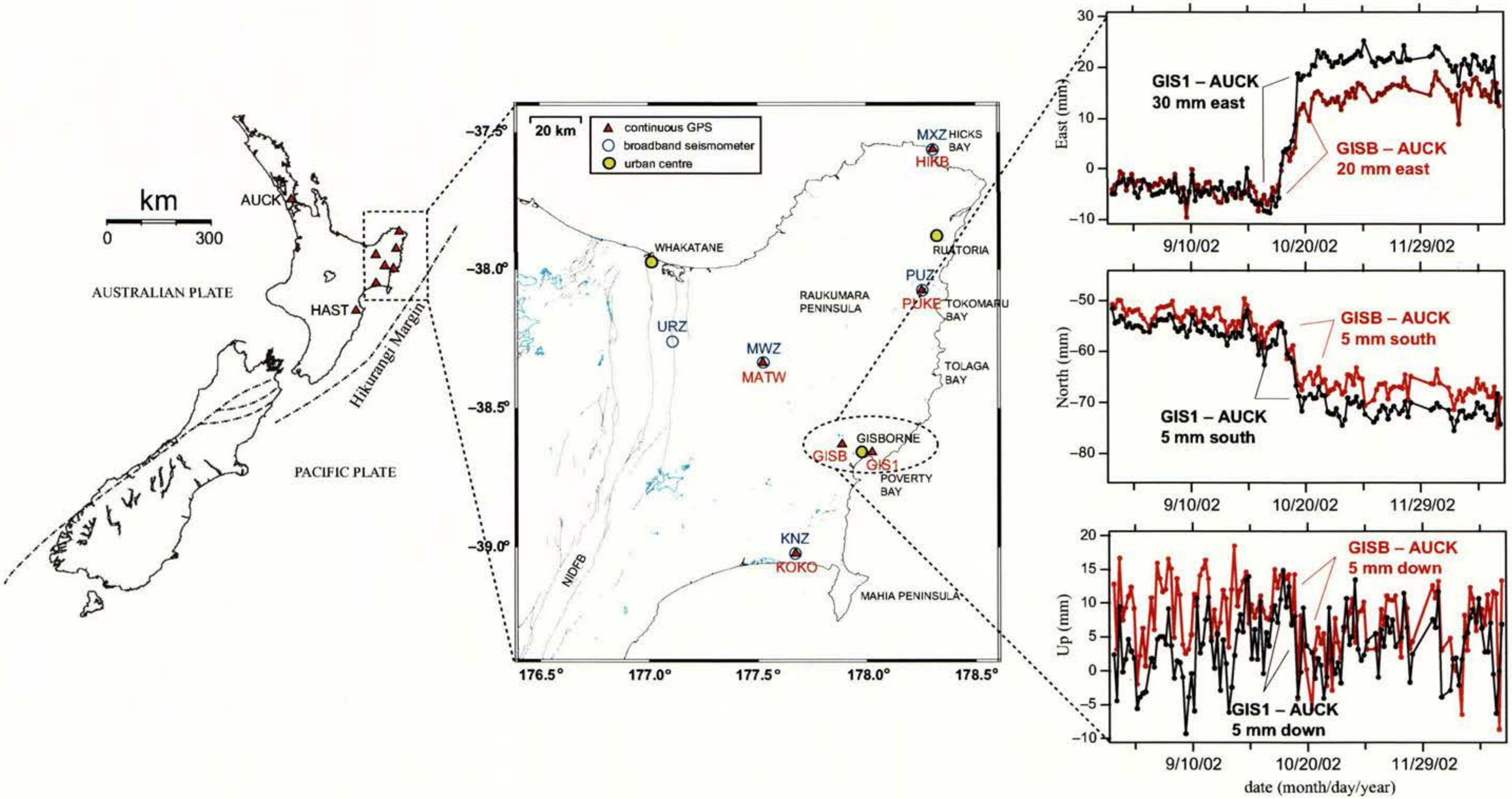

Figure 1.1: **Raukumara Peninsula continuous GPS and broadband seismic station locations.** The Gisborne 2002 slow slip events recorded on GeoNet continuous GPS station GISB and station GIS1. Station position time series are given relative to a continuous station in Auckland (AUCK), which is on the stable Australian plate. The plate boundary between the Pacific and Australian plates is shown schematically as a dashed line.

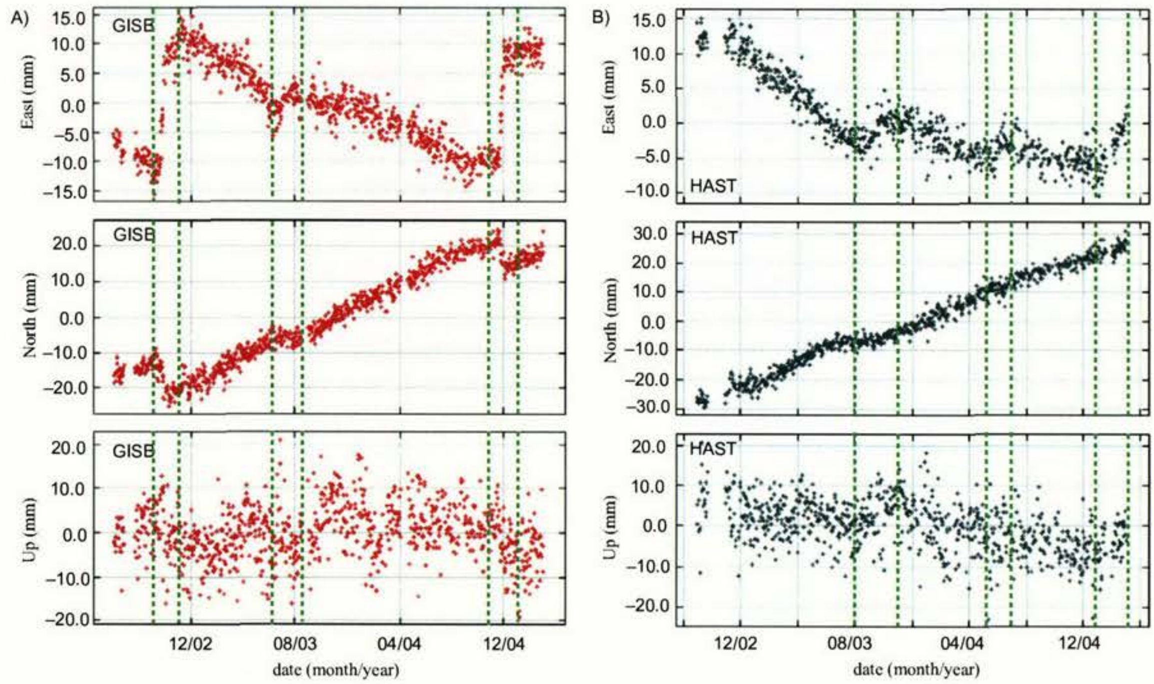

Figure 1.2: **GISB and HAST continuous time series** between 2002 and February 2005. The figure is taken from the GeoNet website and the station position time series are shown relative to the International Reference Frame, ITRF2000. Dashed lines bound slow slip events. GPS station locations are shown are shown in Figure 1.1

and re-evaluate ideas about deformation on the northern Hikurangi margin and Raukumara Peninsula.

As the stations which recorded the Gisborne 2002 slow slip event are close together, and define a line orthogonal to the strike of the subduction zone, the margin parallel extent of the October 2002 slow slip event can not be determined from continuous GPS alone. In an effort to better constrain the spatial extent and slip parameters for the 2002 event we combine recent continuous GPS records with campaign GPS data gathered between 1995 and 2004. The campaign data not only give a much better spatial coverage of the peninsula, but also yield a longer history than the continuous GPS data. This broader spatial and temporal record allows us to examine the frequency characteristics of the slow slip events occurring beneath the Raukumara Peninsula, and the effect such events have on regional deformation patterns.

Since the region's first recorded slow slip event in 2002, the network of continuous GPS and broadband seismometers on the Raukumara Peninsula has been significantly expanded as part of the GeoNet project. GeoNet comprises a national network of continuous GPS, broadband and strong motion seismometers, and volcano-monitoring technologies, with regions such as the Raukumara Peninsula targeted according to their geological and seismic hazard.

Other slow slip events have now been recorded on continuous GPS stations along the Hikurangi margin. In mid 2003 a second slow slip event was recorded at continuous GPS stations in Gisborne (GISB, GIS1), and in Hastings (HAST) to the south (Figure 1.2). This event was much smaller than that observed in 2002, with approximately 5 mm of surface displacement to the southeast recorded at GISB and HAST over 25–30 days. The event was recorded on HAST approximately one month later than at GISB indicating that the slow slip deformation propagated southward along the subduction interface. Two further events, generating similar surface displacements have been recorded since then at HAST (Figure 1.2B).

In November 2004, another slow slip event was recorded at GISB (Figure 1.2A). This event was similar in size to the October 2002 event, with about 20–25 mm of surface displacement recorded at GISB. While there were many more continuous GPS stations in the Gisborne area in 2004 than there had been two years previously, GeoNet stations MATW, PUKE, KOKO (Figure 1.1) had only recently been installed at the time of the event. As the station position time series for these stations are still very short, and systematic annual variations are not accurately known, we cannot yet determine whether they were affected by the November 2004 slow slip event. In general, data have only been acquired from privately run station GIS1 when required to confirm deformation patterns recorded on GISB. This was especially important in 2002 when no other stations were running in the region.

1.3 Tectonic setting of the Raukumara Peninsula

The Raukumara Peninsula is the easternmost part of the North Island, New Zealand (Figure 1.3). Offshore to the east, the Hikurangi Trough marks the present day plate boundary where the Pacific plate subducts beneath the North Island. The crust of the subducting Pacific plate, forming the Hikurangi plateau, is dotted with volcanic edifices and seamounts (e.g. Collot et al., 2001). To the east of the Raukumara Peninsula it has a thickness of about 13 km (e.g. Davey and Wood, 1994). The Pacific plate motion relative to the Australian plate along the northern Hikurangi margin is 45 mm/yr at 266° (DeMets et al., 1994), as shown in Figure 1.3. To the south the plate motion slows and becomes more parallel to the transform plate boundary formed by the Marlborough faults and the Alpine Fault (e.g. DeMets et al., 1990; Anderson et al., 1993).

Seismicity in the Raukumara region is the expression of ongoing subduction along the Hikurangi margin. The majority of earthquakes under the Raukumara Peninsula form a westward-dipping zone which defines the down-going Pacific plate (Ansell and Bannister, 1996). The subduction interface lies at about 15 km depth off the east coast of the Raukumara Peninsula (Figure 1.4; Ansell and Bannister, 1996; Reyners, 1998), and offshore has a dip of less than 10°. Between 15 and 30 km depth below the Raukumara Peninsula the dip of the subduction interface is 15°, and with greater depth the dip increases (Ansell and Bannister, 1996).

To the west, active backarc extension forms the Taupo Volcanic Zone (TVZ, e.g. Reyners, 1980). The Raukumara Peninsula is separated from the TVZ by the the North Island Dextral Fault Belt (NIDFB) (Beanland and Haines, 1998), a zone of right-lateral strike-slip faulting which extends from the southern North Island to the Bay of Plenty.

1.3.1 Crustal structure and neotectonics

The geology of the Raukumara region is depositionally and structurally complex, with rapid changes in facies, common unconformities, and numerous faults and major thrust sheets (e.g. Mazengarb and Speden, 2000). Uplift and erosion rates are high, and few active fault traces are preserved (Mazengarb and Speden, 2000).

Valuable information about the vertical tectonic history of the Raukumara Peninsula during the Holocene is provided by marine terraces (Ota et al., 1992). In a small area of the east coast of the Raukumara Peninsula, 20 km northeast of Gisborne, marine terraces indicate one of the highest known average rates of coastal uplift in New Zealand — approximately 4 mm/yr (Ota et al., 1992). Ota et al. (1992) interpreted these terraces to have formed in six discrete uplift events caused by large local earthquakes over the past ca. 7 kyr. On the east coast of the Raukumara Peninsula, work currently being undertaken by Wilson et al. (2004) on uplifted estuarine sediments provides evidence for tectonic uplift in this region since ca. 6 ka. Berryman et al. (1989) found the likely cause of coastal uplift along the Hikurangi margin to be movement of steep west-dipping reverse faults within the accretionary prism.

High quality marine and seismic reflection lines, exploration wells and seabed dredge samples are available for the continental shelf area of Hawke Bay[2], south of the Raukumara Peninsula. Barnes and Nicol (2004) used these data to identify a thrust zone which underlies the continental shelf; they show that the Australian plate is currently undergoing contraction on northeast–southwest-striking thrust faults and folds.

[2]Hawke Bay is the bay offshore of the geographical region of Hawke's Bay in the eastern North Island (Figure 1.3.

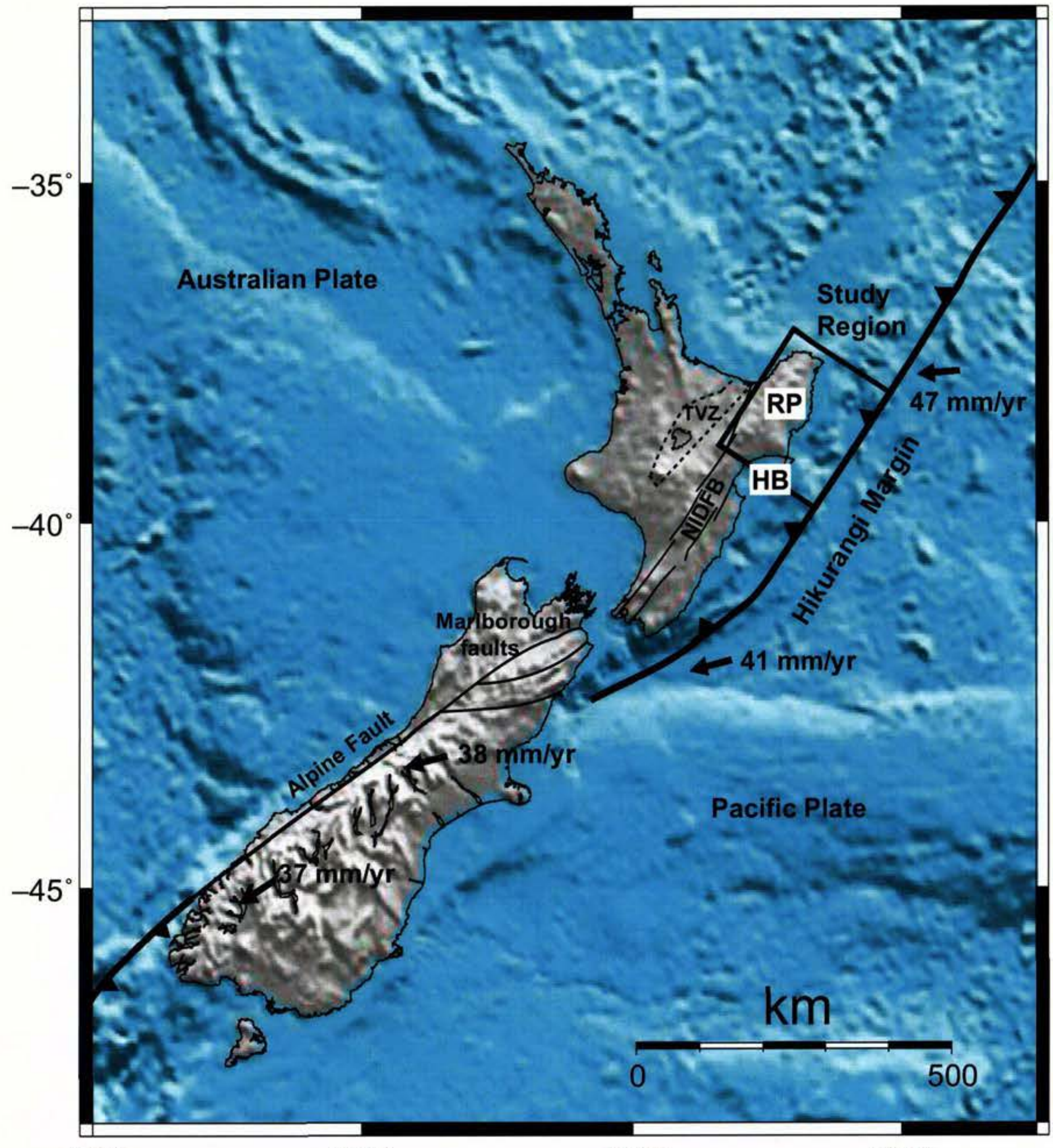

Figure 1.3: **Tectonic setting of New Zealand** showing the Australian-Pacific plate boundary zone. The Raukumara Peninsula (RP) study region is outlined with a box. TVZ — Taupo Volcanic Zone, NIDFB — North Island dextral fault belt, HB — Hawke Bay. Vectors show the motion of the Pacific plate relative to the Australian plate (DeMets et al., 1994).

1.3.2 Geodesy

Geodetic surveys have been undertaken on parts of the Raukumara Peninsula since the 1870s. These were triangulation surveys prior to 1995 when the first GPS campaign was conducted. Such studies, which often link the Raukumara Peninsula to the surrounding regions of the TVZ and Hawke's Bay, have shown strain on the Raukumara Peninsula to be spatially and temporally variable (e.g. Thornley, 1996; Árnadóttir et al., 1999).

Using triangulation data collected in the 1920s and 1976, and GPS data collected in 1995, Árnadóttir et al. (1999) calculated strain rates for the Raukumara Peninsula. They found that, between the 1920s and 1976 the strain rate and orientation over the Raukumara Peninsula was significantly different from that of the 1976 to 1995 epoch. Árnadóttir et al. (1999) suggested this variability could be the result of aseismic[3] slip on the shallow subduction interface occurring sometime between 1926 and 1976. Their study is described in detail in Chapter 4, where we also present strain calculations from this study.

Modelling of geologic data by Beanland and Haines (1998) and geodetic data by Walcott (1984), Beavan and Haines (2002) and Wallace et al. (2004) shows clockwise rotation of the Raukumara Peninsula away from the TVZ. Based on their estimates of the block kinematics Wallace et al. (2004) suggest that this rotation of the eastern North Island occurs because of the southwardly-increasing thickness, and subsequent resistance to subduction, of the Hikurangi Plateau.

Inter-seismic coupling

When inferred from geodetic measurements, the degree of inter-seismic coupling indicates the state of slip across a fault surface in the time between earthquakes (e.g. Wang and Dixon, 2004). A plate boundary that is said to be creeping is slipping at the plate convergence rate. In comparison, a fully locked or coupled plate boundary is not slipping. This lack of slip will in many cases be made up by slip in a subduction earthquake or slow slip event. Inter-seismic coupling can be estimated by geodetic measurements because it causes temporary elastic strain in the earth's crust.

The Raukumara Peninsula offers a good opportunity to study inter-seismic coupling on the northern Hikurangi subduction interface. Here the forearc of the subduction zone occurs on

[3]Árnadóttir et al. (1999) use the term 'aseismic slip' to describe slow slip deformation on the subduction interface; we prefer the term slow slip as it does not preclude the association of slow slip with seismic tremor activity as has been extensively reported in Cascadia and Japan (e.g. Rogers and Dragert, 2003; Obara, 2002).

land and geodetic measurements can be made above the shallow subduction interface. Work using both geodetic (e.g. Darby and Beavan, 2001; Nicol and Beavan, 2003; Wallace et al., 2004) and seismic (Reyners, 1998) data from the length of the Hikurangi subduction margin shows that in general the downdip width of inter-seismic coupling is broad in the south and narrows to the northeast. Higher spatial resolution modelling of GPS velocity vectors by Wallace et al. (2004) highlights a locked patch beneath the Raukumara Peninsula near Tolaga Bay, which we discuss in detail in Chapter 3. This area coincides with the region of high coastal uplift documented by Ota et al. (1992).

1.3.3 Seismicity

The general distribution of seismicity for the Raukumara Peninsula and surrounding regions is shown in Figure 1.4.

Historical earthquakes

The majority of large historical earthquakes in Hawke's Bay region south of the Raukumara Peninsula are associated with strike-slip faulting in the Australian plate (H2, H4 and H5 in Figure 1.4, Doser and Webb, 2003). In comparison, the 1966 Gisborne earthquake (GI in Figure 1.4) of $M_w = 5.6$ had a focal mechanism consistent with slip on the plate interface (Webb and Anderson, 1998), and the 1993 Ormond earthquake (OR in Figure 1.4) of $M_w = 6.2$ occurred within the subducted Pacific plate (Reyners et al., 1998) .

In 1947, two locally generated tsunami affected coastal areas of the Raukumara Peninsula near Gisborne and Tolaga Bay (Downes et al., 2004). The associated earthquakes (TB and P1 in Figure 1.4) were estimated to have $M_w \approx 7.0$ (Doser and Webb, 2003). Downes et al. (2004) reported that the earthquake records were characterised by emergent, low-amplitude P wave arrivals, long rupture duration and a long train of low-frequency surface waves, suggesting that the slip velocity was relatively slow. A slow slip event on the subduction interface offshore of Gisborne may have been causally related to tsunamigenic earthquakes (Árnadóttir et al., 1999).

[4] http://data.geonet.org.nz/QuakeSearch/index.jsp

[5] http://www.seismology.harvard.edu/CMTsearch.html

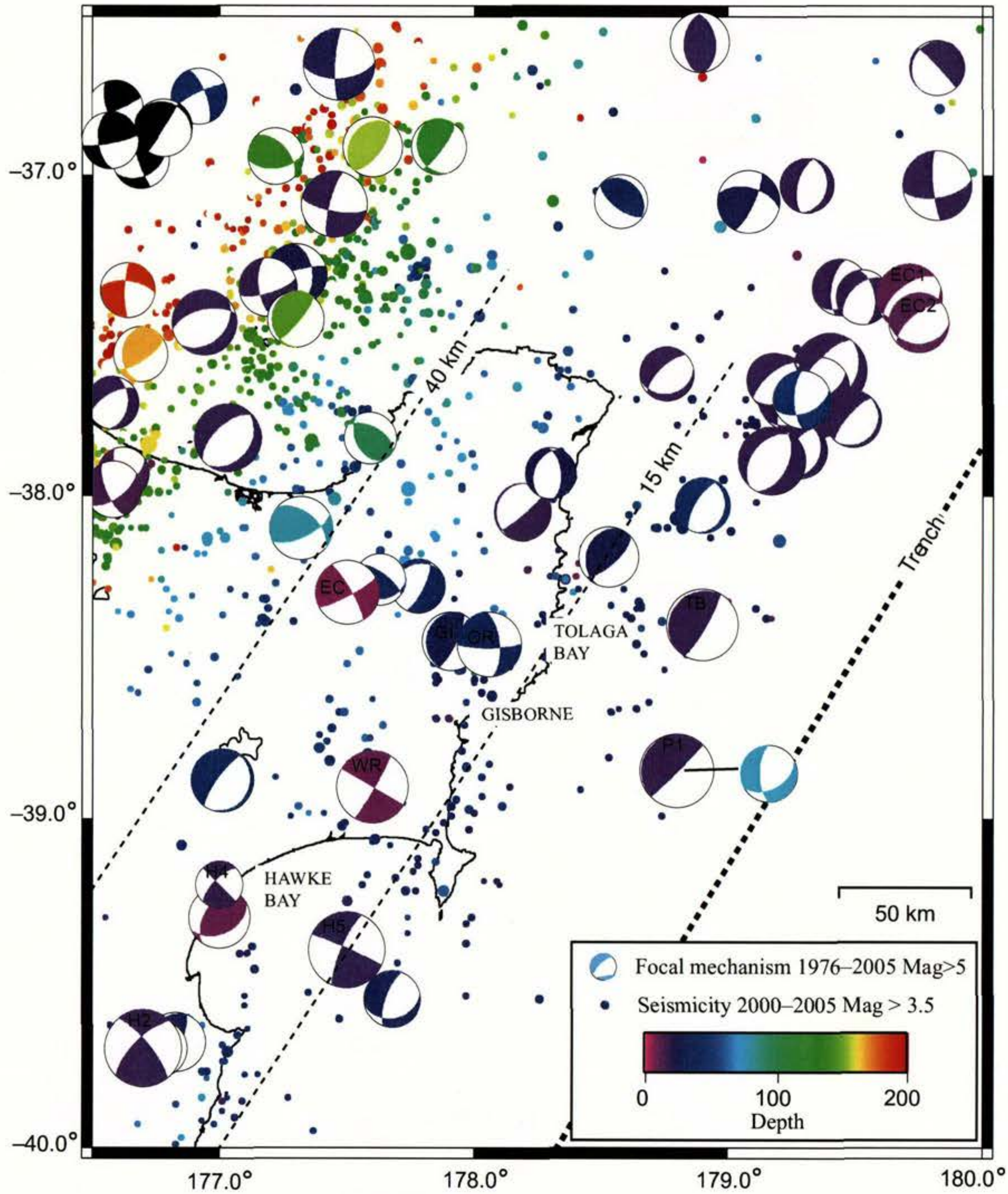

Figure 1.4: **Distribution of seismicity and earthquake focal mechanisms for the Raukumara Peninsula.** Earthquakes larger than magnitude 3.5 occurring between 2000–2005 are shown as dots, colour-coded according to depth. These earthquakes are from the GeoNet catalogue[4]. Depth-restricted earthquakes are not shown. Focal mechanisms show earthquakes larger than magnitude 5; where labelled, they are taken from Doser and Webb (2003). Focal mechanisms for earthquakes deeper than 200 km are coloured black. Unlabelled focal mechanisms are from the Harvard CMT catalogue[5] from 1976 to the present. The approximate depth of the subduction interface beneath the Raukumara Peninsula, after Reyners (1998), and the location of the trench are shown. One focal mechanism is offset from its original location, which is shown with a black dot. EC: East Cape; EC1, EC2: East Cape 1 and 2; TB: Tokomaru Bay; P1: Poverty Bay 1; OR: Ormond; GI: Gisborne; WR: Wairoa; H2, H4, H5: Hawke's Bay.

Seismicity and velocity structure

Using well constrained focal mechanisms determined for 117 events on the Raukumara Peninsula, Reyners and McGinty (1999) found that the distribution of seismicity on the northern Hikurangi margin varies significantly along strike. They found that events shallower than 9.5 km in the overlying Australian plate were predominantly normal. In contrast, events deeper than 9.5 km in the overlying plate were predominantly thrust northeast of Tolaga Bay, compared to predominantly strike slip events in this depth range southwest of Tolaga Bay (Reyners and McGinty, 1999). Southwest of Tolaga Bay thrusting was limited to two events near the plate interface (Reyners and McGinty, 1999).

The observed variation in earthquake mechanisms along-strike of the northern Hikurangi subduction margin corresponds to a change in thickness and velocity structure of the crust of the overlying Australian plate (Reyners, 1998). Northeast of Tolaga Bay the crust is thin ($\sim$18 km thick), while in the southwest it is much thicker ($\sim$36 km, Reyners et al., 1998). In the northeast there is an extensive low velocity zone in the lower crust underlying the Raukumara Range where subducted sediment is interpreted to pond against the stronger mantle, which acts as a backstop (Reyners, 1998; Eberhart-Phillips and Chadwick, 2002). The extensive low-velocity zone dies out southwest of Tolaga Bay, where the thick overlying crust allows sediment to be subducted much deeper along the plate interface (Reyners, 1998; Eberhart-Phillips and Reyners, 1999; Eberhart-Phillips and Chadwick, 2002). The change in thickness of the Australian plate crust corresponds to a change in inter-seismic coupling northeast of Tolaga Bay as suggested by the earthquake focal mechanisms (Reyners and McGinty, 1999) and GPS (Wallace et al., 2004).

1.3.4 Summary

The Raukumara Peninsula is a complex and actively deforming region. Evidence from marine terraces and offshore seismic surveys shows the Raukumara Peninsula to be a region of contraction, with large prehistoric earthquakes and rapid tectonic uplift occurring on steep, west dipping reverse faults (e.g. Barnes and Nicol, 2004; Berryman et al., 1989). Geologic and geodetic studies show that the Raukumara Peninsula is rotating away from the TVZ (e.g. Beavan and Haines, 2002) and that, in general, the zone of inter-seismic coupling on northern Hikurangi subduction interface is shown to be narrow (e.g. Wallace et al., 2004). A locked zone on the plate interface below the Tolaga Bay region (Wallace et al., 2004) coincides with a major change in the style of deformation for the region observed in seismic, geodetic and geologic studies (Eberhart-Phillips and Chadwick, 2002; Ota et al., 1992; Reyners, 1998; Reyners and McGinty, 1999). Subduction and associated deformation along the northern Hikurangi margin is complicated by the presence of seamounts on the subducting Pacific plate (e.g. Collot et al., 2001).

Slow slip events have only recently been recognised in continuous GPS observations from the Raukumara Peninsula (Beavan et al., 2003). This previously unaccounted-for style of deformation is a key factor in the understanding of how inter-plate deformation along the northern Hikurangi subduction margin is accommodated. Studies of slow slip from other subduction zones worldwide help us to understand the broader tectonic context of slow slip events on the northern Hikurangi margin.

1.4 Worldwide occurrence of slow slip events

Here we examine three case studies of slow slip deformation from Pacific Rim subduction zones. These examples, from the Pacific Northwest of North America (Cascadia), Mexico, and Japan, represent key advances in the study of slow slip deformation on subduction zones, using both GPS and seismic data sources. They help establish a context for understanding similar events in New Zealand.

In general, slow slip events from around the Pacific rim are found to occur near the base of the seismogenic zone on the subduction interface (e.g. Dragert et al., 2001; Obara et al., 2004; Larson et al., 2004). The downdip limits for large subduction zone earthquakes are found to correspond to the depth on the subduction thrust where either, the temperature reaches about 350° (the transition to thermally activated stable-sliding behaviour for crustal rocks) or the subduction thrust intersects with the overlying continental forearc Moho and hydration of the forearc mantle rocks enables stable-sliding (e.g. Hyndman et al., 1997; Peacock, 2003; Oleskevich et al., 1999). It is hypothesized that instabilities at the base of the seismogenic zone caused by time varying rheology and friction cause slow slip Dragert et al. (2001).

1.4.1 Cascadia margin

Off the west coast of Canada and the northwestern United States the Juan de Fuca plate is subducted beneath the North American plate. Continuous GPS networks on the coast show an inter-seismic motion of about 9 mm/yr to the northeast (e.g. Dragert et al., 2001). In the past, this segment of the plate boundary has experienced repeated great thrust earthquakes (e.g. Dragert et al., 2001), and current geodetic measurements confirm that the Cascadia subduction margin under British Columbia is currently locked to ca. 25 km depth (e.g. Oleskevich et al., 1999), (ie. no slip is occurring inter-seismically on the part of the subduction boundary that underlies the offshore continental slope) .

In the summer of 1999 a cluster of seven continuous GPS sites in southwest British Columbia and northwest Washington State briefly reversed their direction of motion (Dragert et al., 2001). The recorded transient slip was to the west, in the direction of the North American plate motion relative to the Juan de Fuca plate. Total horizontal displacements recorded during the transient slip event were 2–4 mm over 6–15 days. The deformation migrated to the northwest, parallel to the subduction zone, interpreted as affecting an area 50 km $\times$ 300 km on the subduction interface at between 25 and 42 km depth. Modelling showed that the surface deformation was compatible with 2 cm of slip on the subduction interface downdip of the main seismogenic zone (Dragert et al., 2001). The total cumulative moment of this slow slip event was 1.35×10^{19} N m. No obvious seismic triggers or anomalous seismicity were associated with the event.

Miller et al. (2002) subsequently found that slow slip events on the Cascadia margin beneath British Columbia are episodic, with a repeat interval of 13–16 months. The events are all very similar in size, as can be seen in Figure 1.5, where the inter-seismic trend reflecting coupling between the Juan de Fuca and North American plates is overlain by the steeper sawtooth pattern of the episodic slow slip, which releases some of the accumulated inter-seismic strain.

While not accompanied by earthquakes, the episodic slow slip events recorded on the Cascadia margin under British Columbia were found to correlate with unique seismic signals observed on continuous seismometers in the region (Figure 1.5, Rogers and Dragert, 2003). These seismic signals, which were first observed in Japan (see below, Obara, 2002), are very similar in appearance to tremor observed in volcanic settings due to magma migration, and have been called 'non-volcanic tremor'. The tremor recorded on the Cascadia margin has its energy mainly in the 1–5 Hz range, a lower frequency content than nearby small earthquakes which have most of their energy above 10 Hz (Rogers and Dragert, 2003). When recorded on a single station, the tremor is unremarkable as it falls within the frequency range (1–5 Hz) of environmental and cultural noise. However in Cascadia, as in Japan (Obara, 2002), the tremor was observed simultaneously at several stations across the network, allowing noise sources to

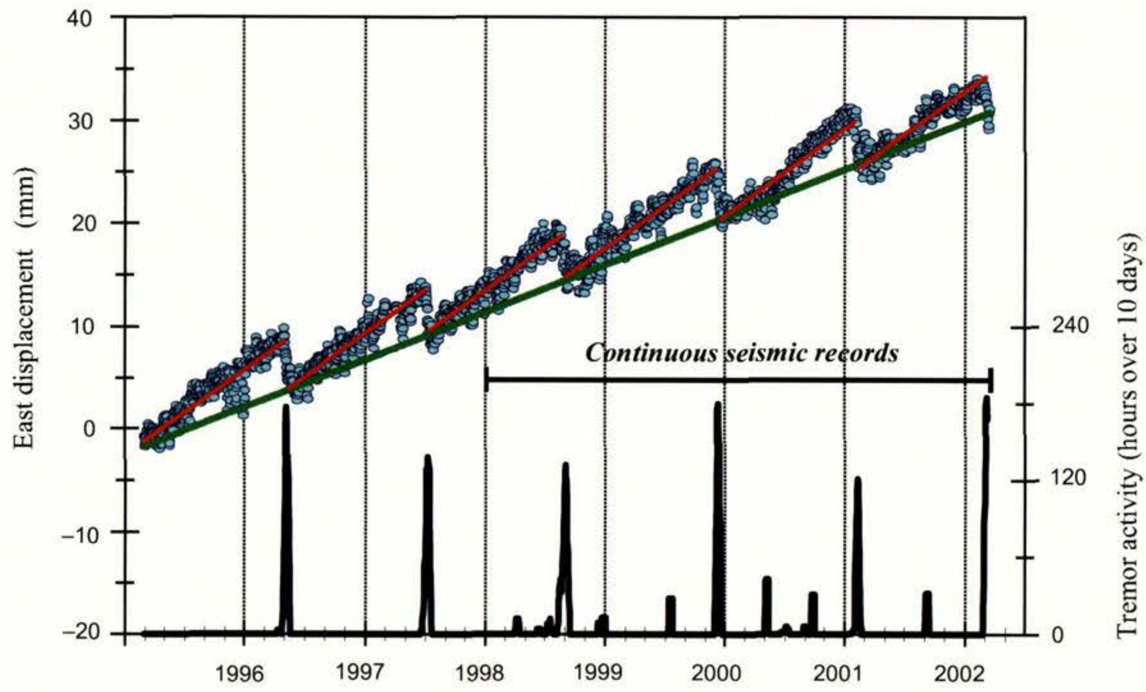

Figure 1.5: **Episodic slow slip from the Cascadia Margin** showing long term trend of plate motion due to the convergence of the Juan de Fuca and North American plates, punctuated by the reversals caused by episodic slow slip. The occurrence of non volcanic tremor is correlated to slow slip events. Figure is taken from Rogers and Dragert (2003).

be discounted. It has an emergent onset, and occurs as pulses of energy for a few minutes to several days. In Cascadia, tremor was found to be strongest on horizontal seismographs and propagated across the seismic network at shear wave velocities. The strongest tremor could be detected as far as 300 km from the source region, and the maximum tremor amplitudes were at least a factor of ten larger than the minimum detectable tremor amplitudes (Rogers and Dragert, 2003).

Other workers studying geodetic and seismic data from the Cascadia margin have observed episodic slow slip and tremor on continuous networks in western Washington, Oregon and northern California, USA (e.g. McCausland et al., 2004; Szeliga et al., 2004). In the south, slow slip events accompanied by tremor have occurred on average every 10.9 ± 1.2 months at least since 1998 (Szeliga et al., 2004). These events result in westward station displacements of up to 4 mm. Tremor bursts were similar in character to the northern Cascadia examples, and occurred in bursts of between 10 mins to over 20 hrs (Szeliga et al., 2004).

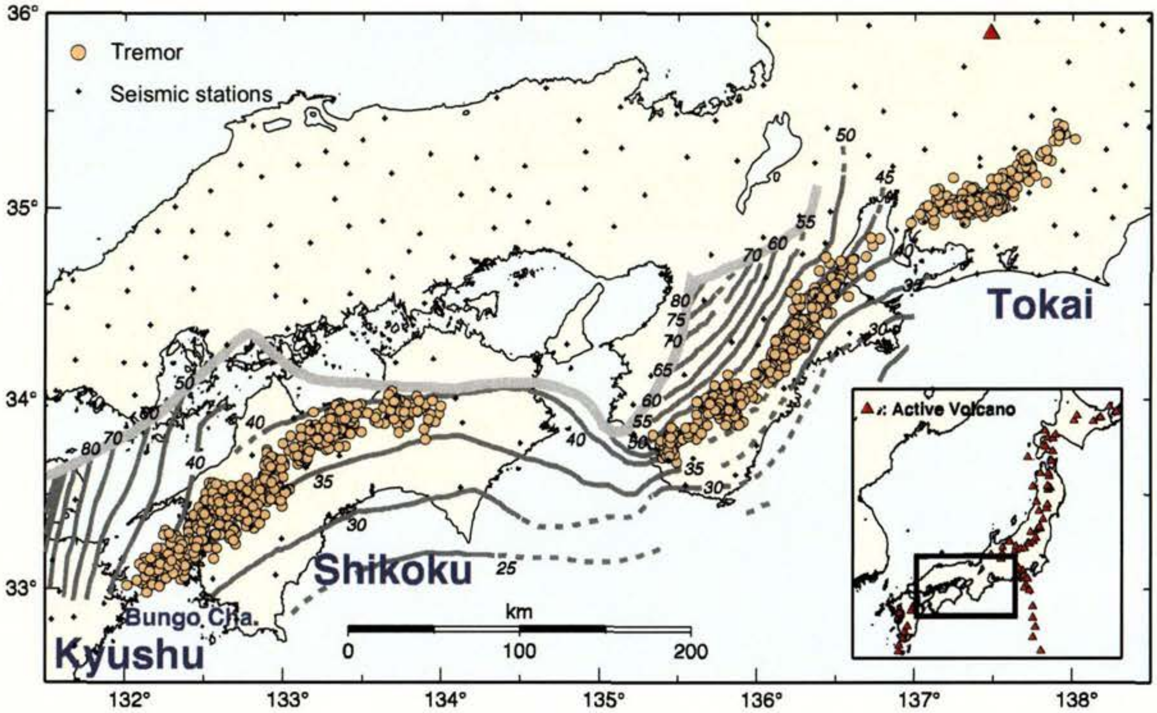

Figure 1.6: **Epicentral distribution of deep long-period tremor in southwest Japan in 2001.** The depth contour lines for the subducting Philippine sea plate are shown. The leading edge of the subducting Philippine sea plate is a thick gray line. Figure is taken from Obara (2002).

1.4.2 Japan

Southwest and central Japan, comprising the islands of Kyushu, Shikoku and the Tokai region (Figure 1.6), lie on the south eastern margin of the Eurasian Plate, where the Philippine Sea Plate is being subducted along the Nankai Trough. To the north, the Pacific plate subducts under the North American plate. Slow slip events recorded in this region vary significantly in duration and frequency. Deep, long-period, non-volcanic tremor has been recognised and located in a belt-like zone from the Bungo Channel to Tokai (Obara, 2002).

Bungo Channel

In 1997 a slow slip event occurred beneath the Bungo Channel (Figure 1.6), between Kyushu and Shikoku, in southwest Japan (Hirose et al., 1999). It lasted for one year with maximum slip rates on the subduction interface of 0.6 m/yr (Miyazaki et al., 2003). Fifteen GPS stations onshore from Bungo Channel detected ground motions of up to 3 cm during 1997 (Ozawa et al., 2001), corresponding to thrust motion on the subduction interface. The event began about 1 month after the second of two magnitude 6.7 earthquakes in Hyuganada to the south. The slow slip nucleated separately from the earthquakes and did not propagate from the source of the earthquakes. Rather, starting at about 30–35 km depth on the subduction interface, it propagated to the southwest and downdip (Miyazaki et al., 2003). The accumulated moment for this slow slip event was 8.1×10^{19} N m (Miyazaki et al., 2003). In July 2003 a similar slow slip event began beneath the Bungo channel, this time with an 8 month duration (Ozawa et al., 2004). Associated with this event, the number of low-frequency earthquakes with epicentral depths around 40 km increased in the western part of Shikoku (Ozawa et al., 2004). The Bungo channel region coincides with a change in strike angle of the Nankai trough, and a lateral transition between strong inter-seismic coupling in the north to weak inter-seismic coupling south of Bungo channel (Hirose et al., 1999).

Shikoku

Episodic slow slip events have recently been recognised through changes on a tiltmeter network in the western Shikoku area (Figure 1.6; Obara et al., 2004). These slow slip events, which have a recurrence interval of approximately six months, occur at the base of the seismogenic zone on the subducting plate boundary and are associated temporally and spatially with seismic tremor (Obara et al., 2004). The displacements expected from slow slip modelling are less than 2 mm on the ground surface, making these events difficult to detect using the continuous GPS network.

Tokai

In the Tokai region (Figure 1.6), the continuous GPS network installed in 1994, together with historical surveys, indicates steady strain accumulation over the region, and the presence of a seismic gap (Ozawa et al., 2002). A long duration slow slip event is ongoing in this region (Obara et al., 2004). The event began in January 2001, and between January 2001 and June 2002 south-eastward motion of around 2 cm at four stations in the western Tokai region was recorded (Ozawa et al., 2002). This was opposite to the steady northwestward motion observed on continuous GPS in the region from 1997–1999. Epicentres of tremor associated with the Tokai slow slip are spatially clustered, at depths averaging 30 km, along the strike of the subducting Philippine sea plate over a length of 600 km (Obara, 2002; Obara et al., 2004). They are characterised by long durations, from hours to weeks and have wide source areas (Obara et al., 2004). The tremor was clearly seen for time windows of 35–50 mins. Similar to Cascadia (e.g. Rogers and Dragert, 2003), envelopes from these signals show gradual rise times rather than that of an earthquake which forms a spike and the envelopes are similar for stations across the network (Obara, 2002). The tremor was found to propagate by s-wave velocity and sometimes seemed to be triggered by medium earthquakes or alternately, culminate in an earthquake (Obara, 2002). Due to the long duration, large source area and mobility of the tremor Obara (2002) interpreted that the tremor was generated by movement of fluid on the subduction zone.

1.4.3 Guerrero Gap, Mexico

Subduction of the Cocos plate under southern Mexico accommodates rapid 5–7 cm/yr convergence, with major earthquakes occurring at 30–100 yr intervals (e.g. Lowry et al., 2001). At the 'Guerrero Gap' near Mexico City there is a deficit of seismic energy release causing the accumulation of elastic strain equivalent to 5 m of relative plate motion since the last major earthquake in 1911 (e.g. Lowry et al., 2005). This zone extends for 120 km parallel to the coast. Continuous and campaign GPS data from the Cocos–North American plate boundary in Guerrero, southern Mexico indicate that four large and four smaller slow slip events (Larson et al., 2004; Lowry et al., 2005) have occurred in the region since 1992. Taken together, these eight events suggest that slow slip in Guerrero recurs at an interval of 1.07 ± 0.05 years. While Guerrero slow slip events do occur regularly, the amount of surface displacement observed during each event varies significantly (Lowry et al., 2005).

In 1998, a slow slip event was recorded on a single continuous GPS station in Guerrero (Lowry et al., 2001). This event generated a displacement in the direction of motion of the North American plate relative to the Cocos plate, of 2 mm east, 26 mm south, 16 mm up over several months. There was no associated seismicity. While displacements at a single site cannot uniquely define a dislocation model, Lowry et al. (2001) combined continuous and campaign GPS data with the well-known location and geometry of the subduction thrust to model the deformation source as slip propagating east to west along the strike of the subduction interface.

Another slow slip event, of different character to the 1998 event, was recorded in Guerrero in October 2001. It occurred over 6–7 months and generated an average slip of 10 cm southwest over an area of the plate interface about 550 km $\times$ 250 km (Kostoglodov et al., 2003). The upper limit of moment release for this slow slip event was about 40×10^{19} N m (Kostoglodov et al., 2003). Two unusual, shallow earthquakes that occurred around the time of the slow slip event may have been causally related to the slow slip event. One, a rare upper plate shallow event, may have been triggered by the initiation of the slow slip or vice versa. Intense aftershock activity lasted for six months and overlapped with the slow slip. In comparison to the 1998 event, the 2001 event occurred on a much larger portion of the subduction interface (Kostoglodov et al., 2003).

Modeling by Larson et al. (2004) shows that strain is accumulating on the subduction interface, and that frictional coupling occurring on part of the subduction interface to $\sim$50 km is released relatively frequently in slow slip events. Above $\sim$25 km depth slow slip events apparently do not relieve a significant fraction of the slip deficit (Larson et al., 2004). Locations and sizes of events are only partially constrained but suggest slow slip is centred downdip of the seismogenic portion of the plate bounding thrust (Larson et al., 2004).

There is currently no continuously recording broadband seismic network in the Guerrero region and to date no tremor has been observed there in association with slow slip (Lowry et al., 2005).

1.4.4 Summary

Slow slip events recorded around the Pacific rim are diverse in duration and frequency. Slow slip events from the Cascadia margin are episodic, predictable and consistent in the size of surface displacement recorded. Their surface displacements are small compared with other examples of slow slip from the Pacific rim, including the Hikurangi subduction margin. In Japan, the duration, frequency and size of events varies significantly; the Shikoku episodic slow slip events are similar in character to those recorded on the Cascadia margin. Slow slip events in the Guerrero Gap are episodic but vary greatly in the size of displacement.

There are also similarities between the Pacific rim slow slip events. In particular, they occur at the base of the seismogenic zone on the subduction interface, although this depth varies between subduction zones and is shallow beneath the Raukumara Peninsula. Examples from Cascadia, Japan and Guerrero illustrate that slow slip events accommodate significant moment release during slip, but the subduction margins are shown to be accumulating strain despite episodic slow slip. Slow slip events occurring in the Bungo Channel region coincide with a change in along-strike coupling on the subduction interface of the Philippine sea plate. Similarly the Gisborne slow slip events coincide with an along-strike change in coupling on the Hikurangi subduction margin. In Cascadia and Japan seismic tremor has been recognised in association with slow slip events.

In the context of slow slip on the northern Hikurangi margin these observations from Pacific rim slow slip events provide a useful starting point for our analyses of slow slip events under the Raukumara Peninsula. Studies from the Guerrero Gap illustrate the constraints of working with smaller continuous GPS networks, but also the potential of studying campaign time series to identify slow slip events that occurred prior to installation of continuous networks (e.g. Larson et al., 2004). Studies of seismic tremor associated with episodic slow slip in Cascadia and Japan provide a framework for similar studies on the northern Hikurangi margin (e.g. Rogers and Dragert, 2003; Obara, 2002). Slow slip events are shown to be an important contributor to moment release at subduction plate boundaries (e.g. Kostoglodov et al., 2003; Dragert et al., 2001; Miyazaki et al., 2003). However, the persistence of strain accumulation on subduction zones where slow slip is occurring has important implications for assessing the seismic hazard on the northern Hikurangi margin (e.g. Larson et al., 2004).

1.5 Project overview

In this introduction, we have described slow slip events recorded on the Raukumara Peninsula since 2002 using continuous GPS and established a regional and global context for those slow slip events. In the rest of this thesis, we seek to constrain the frequency and magnitude characteristics of slow slip events on the northern Hikurangi subduction margin using campaign GPS data spanning ten years. To achieve this objective we:

1. Re-process existing GPS data collected during campaigns in 1995, 1997 and 2001, together with new campaign GPS data collected in 2004, using modern software and uniform satellite orbits (Chapter 2).

2. Interpret the combined 1995–1997–2001–2004 data with knowledge of other slow slip events recorded using continuous GPS on the northern Hikurangi subduction margin. We predict the time between events with surface displacements similar in size to the 2002 Gisborne slow slip event and modelled the slip region for that event. We also investigate how the distribution of coupling on the northern Hikurangi subduction margin evolves with time (Chapter 3).

3. Examine what effect the Gisborne 2002 slow slip event had on regional deformation patterns, including strain, using the improved spatial coverage of the campaign GPS data set over the Raukumara Peninsula (Chapter 4).

4. Make a preliminary study into the possible association of seismic tremor with the slow slip events recorded on the Raukumara Peninsula based on methods from Japanese and North American scientists (Chapter 5).

Chapter 2

Global Positioning System (GPS) data

Prior to 1995, geodetic observations of deformation for the Raukumara Peninsula and Gisborne area were obtained during two triangulation surveys undertaken in 1924–1926 and 1976–1977 respectively. These two surveys provide a foundation of geodetic information subsequently built upon by ten years of regional GPS observations. Since 1995, campaign-style GPS surveys have been conducted four times in the Raukumara region, most recently in 2004 in response to the Gisborne 2002 slow slip event. The campaign data offer a broader perspective on the slow slip event than obtainable with limited continuous GPS data alone. Combined with recent continuous GPS records (Figure 2.1; Table 2.1), the campaign surveys provide improved spatial control for earth deformation studies in the Raukumara region, and the extended data record allows temporal change to be examined. In order that comparisons can be made between the four campaign data sets, we have re-processed all the campaign data, together with continuous GPS data. To ensure consistency and compatibility between the results, , we use identical processing techniques and uniform precise satellite orbital information for all epochs. The data collection and processing are described here.

2.1 Raukumara Peninsula GPS campaign surveys

We use data from the four regional GPS campaigns, that have been carried out on the Raukumara Peninsula in the past ten years. Details of each campaign are given here.

- **1995:** GPS measurements of a network of 92 stations in the Raukumara region were made in January 1995 by teams from Victoria University of Wellington (VUW) and the New Zealand Department of Survey and Land Information (DOSLI; Árnadóttir et al., 1999). Trimble 4000SSE receivers were deployed by VUW and collected data for 12 to 48 hours at 39 sites. In addition, DOSLI deployed Leica SR299 dual-frequency receivers, collecting data for up to 8 hours at 56 sites.

- **1997:** In February 1997, 28 sites on the Raukumara Peninsula were surveyed by VUW in conjunction with GNS. Twenty of the sites chosen had been surveyed in 1995. Trimble 4000SSE and TurboRogue receivers were deployed for periods of up to 24 hours.

- **2001:** During a two week period in January 2001, GNS repeated campaign observations at 56 sites in the Raukumara Peninsula network, including some previously unsurveyed sites (Matheson, 2001). This survey was designed to repeat the 1995 and 1997 GPS and earlier triangulation surveys of the Raukumara Peninsula. The 2001 survey also extended into the adjacent Hawke's Bay and Rotorua networks, and increased the sampling density within the Raukumara network. Trimble 4000SSE and SSi, and Ashtech Z-XII receivers were deployed for 24 hours or more at each site.

- **2004:** In March 2004, GNS conducted a campaign in response to the slow slip event recorded on continuous GPS stations near Gisborne in October 2002. We reoccupied a subsection of sites within the Raukumara network, focusing in particular on sites close to Gisborne that had been surveyed frequently in the past. We successfully surveyed all sites with three previous occupations, as well as sites at other strategic locations. Sites A3Q5 and 1297 on the western side of the Raukumara Peninsula were surveyed by a GNS team three weeks after the main campaign to replicate the spatial distribution of the 1995 and 2001 GPS campaigns. In total, 23 sites were surveyed. We used Trimble 5700, Trimble 4000SSE and 4000SSi, and Ashtech Z-XII receivers, and deployed the receivers for a minimum of 48 hours. Appendix A describes our standard field methods for the 2004 Raukumara GPS campaign, and further information is found in Palmer (2004).

During the 1995, 1997 and 2001 surveys, local site A7WE was surveyed continuously, in order that it could be used as a control station. With continuous GPS station GISB running in 2004, this was not necessary and we reoccupied A7WE for 48 hours only.

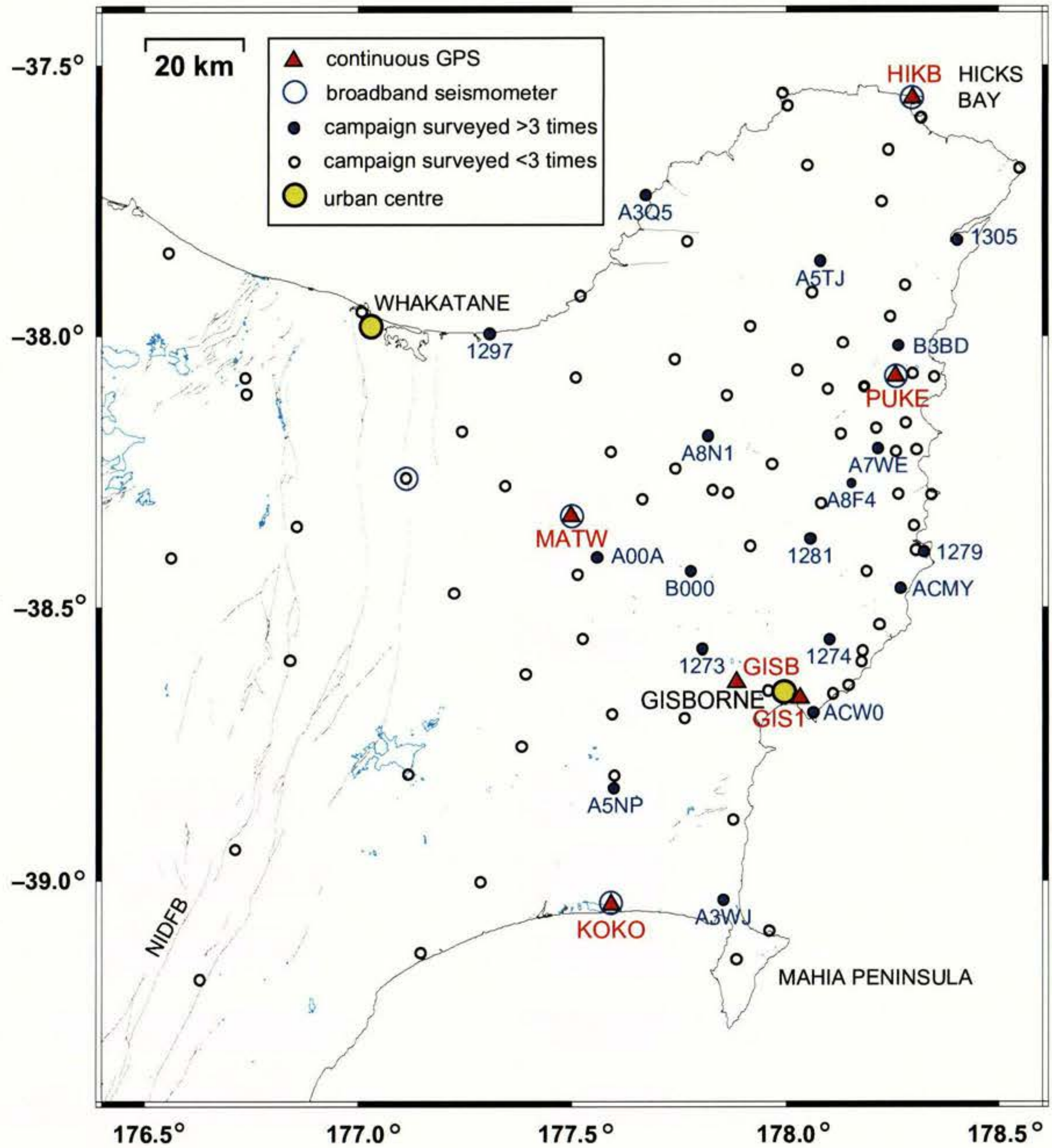

Figure 2.1: **Continuous and campaign GPS sites observed between 1995 and 2004**. Continuous stations are labelled red. Campaign sites which were observed three or more times between 1995 and 2004 are blue. Unlabelled sites have been observed less than three times. NIDFB: North Island dextral fault belt; information on faults comes from the GNS active faults database (1998). Lakes are outlined in blue.

Station	Lat	Long	93	95	96	97	01	03	04	Cont.
1273	-38.575	177.805	×	×	×	×	×	×	×	
1274	-38.558	178.102		×			×		×	
1279	-38.397	178.325		×		×	×		×	
1281	-38.372	178.057		×		×	×		×	
1297	-37.995	177.308		×		×	×		×	
1305	-37.825	178.407	×	×	×	×	×	×	×	
A00A	-38.407	177.559		×		×	×		×	
A3Q5	-37.739	177.672		×		×	×		×	
A3WJ	-38.034	177.855		×			×		×	
A5NP	-38.830	177.597		×			×		×	
A5TJ	-37.861	178.081		×			×		×	
A7WE	-38.206	178.217		×		×	×		×	
A8F4	-38.270	178.155		×		×	×		×	
A8N1	-38.184	177.819		×		×	×		×	
ACMY	-38.464	178.270		×		×	×		×	
ACW0	-38.691	178.064		×		×	×		×	
B000	-38.432	177.778		×			×		×	
B3BD	-38.017	178.266		×		×	×		×	
GIS1	-38.663	178.025								11/2001
GISB	-38.635	177.886								07/2002
HAST	-39.617	176.717								09/2002
HIKB*	-37.561	178.303								05/2003
KOKO*	-39.021	177.674								10/2004
MATW*	-38.336	177.528								04/2004
PUKE*	-38.073	178.257								02/2004

Table 2.1: **Raukumara campaign GPS stations observed on more than two occasions between 1995 and 2004**, shown in (Figure 2.1). Survey epochs were January 1995, February 1997, January 2001 and March 2004. Two first order trig stations, 1273 and 1305, were observed more frequently. An additional 33 stations were observed at two epochs, and 38 stations were observed at only one epoch. The start date (month/year) of continuous operation is shown for each continuous GPS station. HAST is shown in (Figure 1.1). An asterisk indicates continuous GPS sites installed together with continuous broadband seismometers.

2.2 Regional continuous GPS

The first continuous GPS site installed by GNS on the northeastern Hikurangi Margin, GISB, was established at a site west of Gisborne city in February 2002 (Figure 2.1; Table 2.1). A second receiver was installed at Hastings (HAST) in March of the same year. These stations began continuously recording data later in 2002, GISB in July and HAST in late September. HIKB continuous station was installed at Hicks Bay in May 2002. In 2004 the GeoNet continuous GPS network on the Raukumara Peninsula was significantly expanded; stations PUKE and MATW were installed together with broadband seismometers on the central Raukumara Peninsula in February and April 2004 respectively. Most recently KOKO was installed in November 2004 at the KNZ broadband seismometer site near Mahia Peninsula. Data are telemetered continuously to data management centres at Gracefield, Lower Hutt, and Wairakei. These near-real-time data are publicly available via ftp from GeoNet[1].

2.3 GPS data acquistion and processing

The theory of GPS data acquisition and processing is described at length in, for example, Strang and Borre (1997) and Misra and Enge (2001). In addition, succinct explanations of GPS in the context of data processing are given by King and Bock (2002) and Beutler et al. (2001).

2.3.1 General theory

The complete GPS constellation comprises 24 satellites and provides continuous global coverage of between four and eight satellites above 15° from the horizon at any point on the Earth's surface. The satellites orbit at about 20,200 km above the surface of the earth, circling the globe twice every sidereal day (once every 11 hours and 58 minutes) and provide a platform for radio transmitters, atomic clocks and computers (e.g. Beutler et al., 2001; Strang and Borre, 1997; Misra and Enge, 2001).

Each GPS satellite has a very precisely known orbit and broadcasts a message allowing users on earth to recognise the satellite and determine its spatial position. The GPS satellites transmit signals on two carrier frequencies (f_1 and f_2), corresponding to signals L_1 ($\lambda_1 \approx 19$ cm) and L_2 ($\lambda_2 \approx 24$ cm). These signals are modulated by unique codes and a navigation mes-

[1] ftp://ftp.geonet.org.nz/gps

sage to transmit information such as orbital parameters and readings from satellite clocks (e.g. Beutler et al., 2001). All signals transmitted by the satellite are derived from the fundamental frequency of the satellite oscillator.

The navigation message contains information about the satellite clock and satellite orbit. The satellite ephemerides, or orbits, are modulated by gravitational pull from the moon and sun, and by the pressure of solar radiation on the satellites. The navigation message contains updated orbital information based on ground-based observations. The user must decide whether to use these broadcast orbits, which are available in real-time and are accurate to a sub-metre level, or final precise orbits, which are available from the International GPS Service (IGS) after a delay of about ten days. The precise orbits are accurate to several centimeters and contain the highly accurate satellite clock corrections required for processing (e.g. Beutler et al., 2001). IGS operates a global system of satellite tracking stations, data and analysis centers and the resulting data are available online[2].

For global positioning, ground based GPS receivers collect and process signals broadcast from each visible satellite. The preliminary position of the receiver is based on pseudorange measurements from the satellite. The pseudorange is the travel time from each satellite to each receiver, not accounting for clock errors (e.g. Strang and Borre, 1997). This measurement gives the receiver position at an accuracy of several metres. A more accurate position can be calculated from the phase observation, which is the difference in phase of the incoming satellite signal and a receiver generated signal of the same frequency (e.g. Strang and Borre, 1997). When the receiver first begins tracking the satellite, only the fractional part of the phase difference is measured. The initial integer number of cycles between the receiver and the satellite is unknown; this is called the phase ambiguity. The phase ambiguity remains constant for subsequent measurements to the satellite provided the receiver remains locked on the satellite signal. If lock is lost, the phase measurement jumps an integer number of cycles. Such cycle slips are caused by obstruction of the satellite signal or low signal to noise ratio due, for example, to low satellite elevation or failure in the receiver software (e.g. Beutler et al., 2001). Phase ambiguities and cycle slips must be resolved during subsequent GPS processing for highest accuracy results.

[2]http://igscb.jpl.nasa.gov/

2.3.2 Processing strategy

High accuracy in GPS processing is achieved by having a network of receivers making repeated observations, so that differences in the sites' positions can be calculated, rather than their absolute positions (e.g. Strang and Borre, 1997). Differencing receiver and satellite observations can eliminate important error sources in GPS data processing such as receiver and satellite clock errors. The effects of the ionosphere on the phase measurements are eliminated through the formation of linear combinations of the L_1 and L_2 signals.

The primary GPS observation is a one-way difference between a receiver and a satellite (e.g. Strang and Borre, 1997). Next the differences between a pair of receivers observing the same satellite at the same time is taken. This *single difference* eliminates errors common to the satellite, namely the effect of bias or instability in the satellite clock. Finally, the difference between two receivers and two satellites is taken to eliminate errors common to both receivers (e.g. Strang and Borre, 1997). This *double difference* cancels the effects of variations in the receivers' clocks. In a local network, the distance between the receivers used in the double difference is small compared to the 20,200 km-high orbit of the satellite. The signals to each receiver will therefore travel along very similar paths and any path-dependent errors due to the atmosphere or satellite orbit will be common and approximately cancel during double differencing. Baseline vectors and ambiguities jointly estimated in the double difference generation are used in least squares estimation of point coordinates (Strang and Borre, 1997).

Linear combinations of the L_1 and L_2 signals are used in GPS processing to eliminate error sources and resolve phase ambiguities (e.g. Misra and Enge, 2001; Beutler et al., 2001). Ionospheric delay can be eliminated by using the ionosphere-free linear combination:

$$L_3 = \frac{1}{f_1^2 - f_2^2}(f_1^2 L_1 - f_2^2 L_2). \tag{2.1}$$

When resolving phase ambiguities and correcting for cycle slips, the uncertainty in integer estimation depends upon the carrier wavelength. A longer wavelength makes it easier to estimate the ambiguities (e.g. Misra and Enge, 2001). The wide-lane linear combination of L_1 ($\lambda_1 \approx 19$ cm) and L_2 ($\lambda_2 \approx 24$ cm)

$$L_5 = \frac{1}{f_1 - f_2}(f_1 L_1 - f_2 L_2), \tag{2.2}$$

creates a significantly longer wavelength signal of approximately 80 cm. Narrow lane ambiguity resolution uses the L_1 phase data only.

2.4 Processing of the Raukumara GPS data

Following the March 2004 GPS campaign, four campaign data sets spanning nearly ten years were available to us. We processed the complete set of campaign data using two different software packages — GAMIT-GLOBK (Herring, 2001; King and Bock, 2002) and BERNESE 4.2 (Beutler et al., 2001). In both cases RINEX (Receiver INdependent EXchange) format data from the GNS archives were used. We also used data distributed by IGS from a subset of global continuous GPS sites. To ensure consistency, the coordinates and velocities of the global GPS sites in the international terrestrial reference frame, ITRF2000, were used in the processing of all campaign data sets.

Through processing the data with two different software packages we gained familiarity with the input files required, and the outcomes and implications of the different styles of processing. We describe both methods here. Comparisons of the station position time series and their residuals (Figures 2.4, 2.6, 2.5 and 2.7) show that the residual deformations are similar within uncertainties for the two different processing methods. In both cases, we use the standard processing methods as outlined in Herring (2001), King and Bock (2002) and Beutler et al. (2001).

Prior to processing, we organised all the campaign station data into 24 hour blocks according to Coordinated Universal Time (UTC), the international time standard. Processing the data in 24 hr sessions limits the number of parameters which must be solved for. In all but the 1995 Raukumara campaign, GPS data collection was synchronised with the UTC day. For the 1995 data, files were concatenated to align with UTC and to maximise the length of data for each site.

2.4.1 GAMIT-GLOBK processing

The 1995, 1997 and 2001 data were first processed using GAMIT-GLOBK software. This processing was done prior to the 2004 field campaign, and the time series generated were used to review our site selection for the 2004 campaign. Later, the 2004 data set was also processed using GAMIT-GLOBK. The steps for GAMIT-GLOBK processing are shown in Figure 2.2.

GAMIT

Processing in GAMIT is modular. The modules require six main types of input (King and Bock, 2002):

1. station coordinates (good a priori coordinates are required to generate a robust solution);

2. receiver and antenna information for each site;

3. raw phase and pseudorange data contained in RINEX files;

4. satellite list and initial conditions for the satellite orbits;

5. satellite and station clock values;

6. lunar and solar ephemerides, earth rotation information, geodetic datums, and spacecraft and instrumentation information.

In addition to campaign data from the Raukumara Peninsula, we also included data from selected IGS continuous GPS sites in New Zealand, Australia, and the Pacific (Table 2.2).

The programme estimates station coordinates and their covariances, together with the zenith delay of the atmosphere for each station, and satellite orbital and earth orientation parameters. One of the main differences between the GAMIT and BERNESE approaches is that in GAMIT the orbital parameters are estimated in the processing rather than fixed to values estimated by IGS.

All of the receiver and antenna information specific to a particular GPS site occupation is recorded in a station information file. This input file must be created manually. The most important entries are the antenna type, height, and the way in which the antenna height was measured in the field, which directly affects the estimated height in the analysis (King and Bock, 2002). We used information from campaign log sheets to double-check this information prior to processing.

Once all the required files are assembled, a pre-processing programme named MAKEEXP is run to generate additional files required in processing, like navigation and satellite and station

	Site						
Australia	mac1	hob2	(tidb	or tid2)	yar1		
New Zealand (all years)	auck	chat	hoki	mqzg	ousd	(well	or wgtn)
New Zealand (2004)	dnvk	gisb	hast	hikb	mast	taup	trng
Pacific	niuc	noum	kokb	kwj1			

Table 2.2: **Selected IGS continuous GPS sites used for GAMIT processing** for the GAMIT processing of Raukumara Peninsula GPS campaign data. Additional New Zealand stations available only for the 2004 processing are noted.

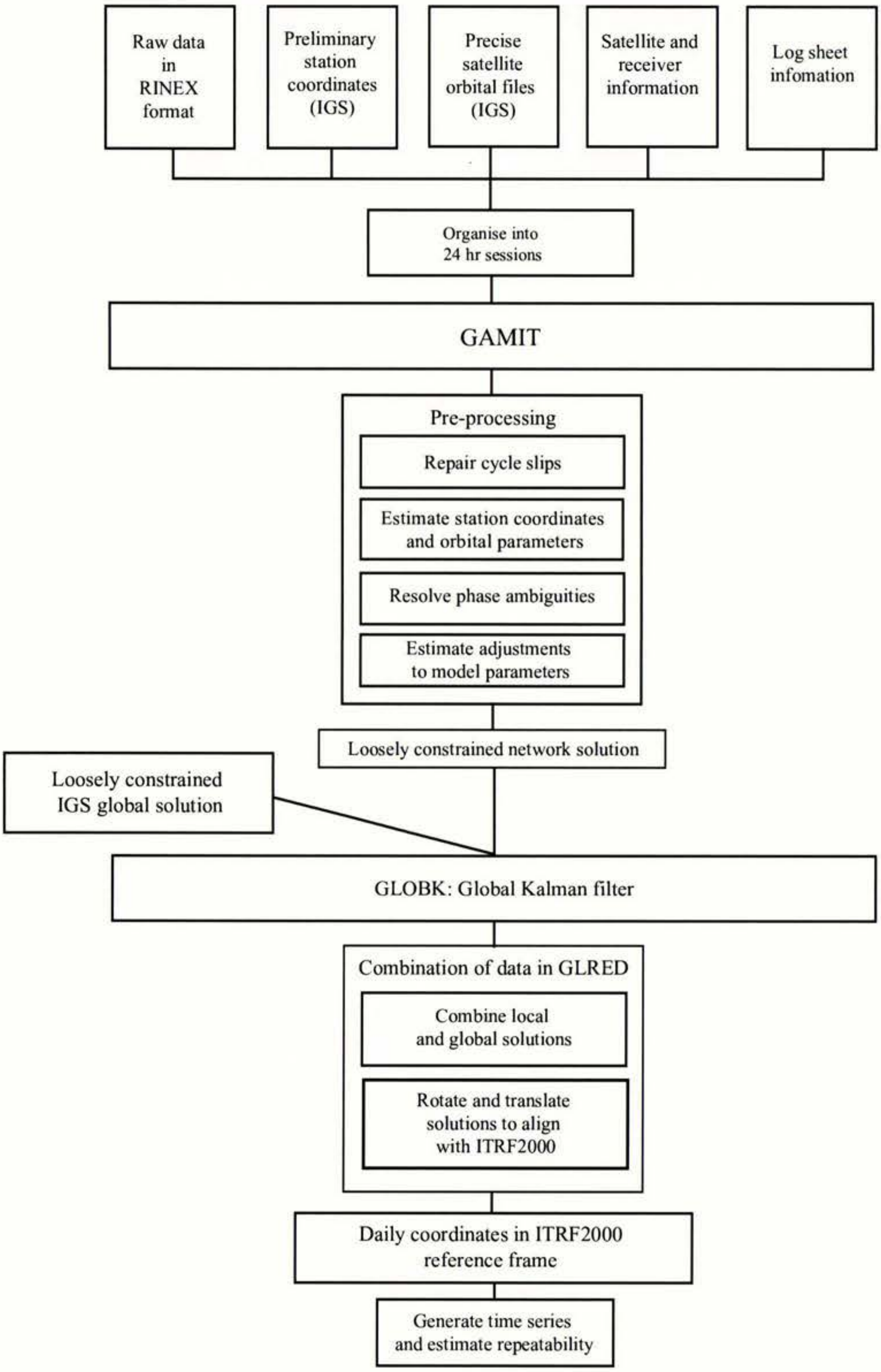

Figure 2.2: **Flowchart of GAMIT-GLOBK processing**

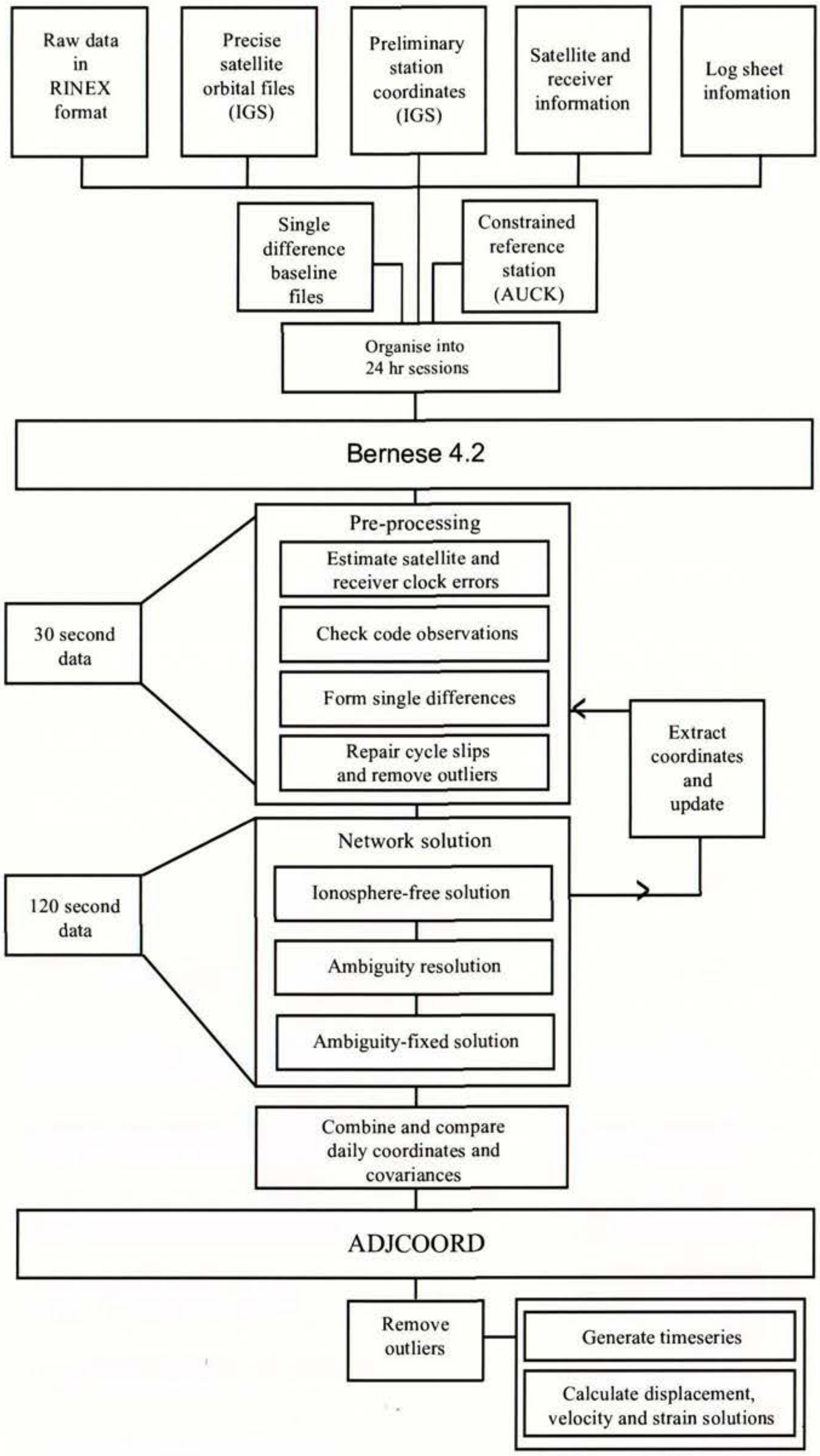

Figure 2.3: **Flowchart of BERNESE-ADJCOORD processing**

clock information files (King and Bock, 2002). MAKEEXP also determines which stations and satellites are to be included in each session.

We completed the main processing in GAMIT as a batch in which individual GAMIT modules were run sequentially with the necessary input files. The three main modules implemented in this process are MODEL, AUTCLN and SOLVE. MODEL creates a file containing prefit residuals and partial derivatives for the observations. These are used in SOLVE to estimate adjustments to the model parameters (King and Bock, 2002). AUTCLN searches for cycle slips and outliers, and creates a file with corrected phase data and prefit residuals. SOLVE performs a least-squares estimate of station coordinates and orbital parameters, and attempts to resolve phase ambiguities for regional stations using wide-lane and narrow-lane resolution (King and Bock, 2002). These steps are repeated iteratively.

The normalised rms (nrms) of the GAMIT solution is a good indicator of whether the data fit the model given the noise level (King and Bock, 2002). A good solution usually produces an nrms of about 0.25; poor solutions (nrms over 0.5) can result, for example, when not all cycle slips have been removed (King and Bock, 2002). For the Raukumara campaign data processing, the posterior nrms for each daily solution was generally less than 0.2.

GLOBK

The main output from GAMIT is a loosely constrained solution (H-) file of parameter estimates (site positions, orbits, zenith delays, earth orientation parameters) and covariances which can be utilized in GLOBK, Global Kalman filter. GLOBK assumes a linear model; it cannot correct deficiencies in the primary analysis due to cycle slips, bad data and atmospheric delay modelling errors, and cannot resolve phase ambiguities (Herring, 2001). These steps must be performed previously in GAMIT.

In order to produce a velocity solution, GLOBK is run iteratively through many days of data. As we were primarily interested in generating time series for analysis, not velocity solutions, we did not use the complete Kalman filter approach in our data processing. Instead we used GLRED, a front end programme for GLOBK, and processed the data from each individual campaign day separately.

The main difference between using GLRED and the complete GLOBK Kalman filter approach is that in GLRED the coordinates of the reference frame for each observation session must be defined, in comparison to GLOBK where the reference frame is defined while obtain-

Site									
Site	algo	onsa	kosg	wtzr	yell	fair	kokb	yar1	sant	tidb
	kour	brmu	wes2	gode	tid2	lhas	irkt	shao	kwj1	hart
	nlib	mdo1	pie1	drao	hob2	tskb	pert	kit3	mali	
	graz	nyal	mets	vill	mas1	braz	ohig	zwen	mate	

Table 2.3: **IGS sites used for global reference frame stabilisation** for the GAMIT-GLOBK processing of Raukumara Peninsula GPS campaign data.

ing the velocity solution (McClusky, MIT website)[3]. The GLRED approach is straightforward for a globally defined reference frame using IGS sites and ITRF2000 coordinates. A loose combination of all the data is produced, then a rotation and translation or scale is applied. Results between the two processes should be very similar (McClusky, MIT website).

For the Raukumara campaign data set we combined the local H-files, from the GAMIT solution, with global H-files, which we downloaded for the days of interest from Scripps Institute of Oceanography Orbit and Permanent Array Center (SOPAC[4]). The combined H-files were used to estimate station positions in the ITRF2000 reference frame for each day and we generated station position time series for individual campaign sites (Figures 2.4A and 2.6A). We placed tight constraints on the ITRF2000 coordinates of selected continuous GPS sites (Table 2.3). The reference frame is defined by these core IGS sites. The global and New Zealand-based solutions are rotated and translated together into the ITRF2000 reference frame.

The uncertainties in daily station positions calculated in GAMIT-GLOBK are nominally 1σ. However, as we did not scale the covariance matrices, and the χ^2 values for each daily solution were consistently around 0.5, these uncertainties represent the 90%–95% confidence level. We also calculated residuals from the station position time series by removing the linear trend of the data (Figures 2.5A and 2.7A).

2.4.2 BERNESE processing

Following the 2004 field campaign, data from all four Raukumara GPS campaigns were also processed using BERNESE 4.2 software. The station coordinate and covariance files generated using the BERNESE processing method were later transferred into GNS software package ADJCOORD. ADJCOORD (Crook, 1992) is a FORTRAN programme for survey adjustment and deformation modelling. It is used to check the data for outliers, which may be removed from the solution. The steps for BERNESE processing are shown in Figure 2.3.

[3]http://www-gpsg.mit.edu/~simon/gtgk/

[4]http://sopac.ucsd.edu/

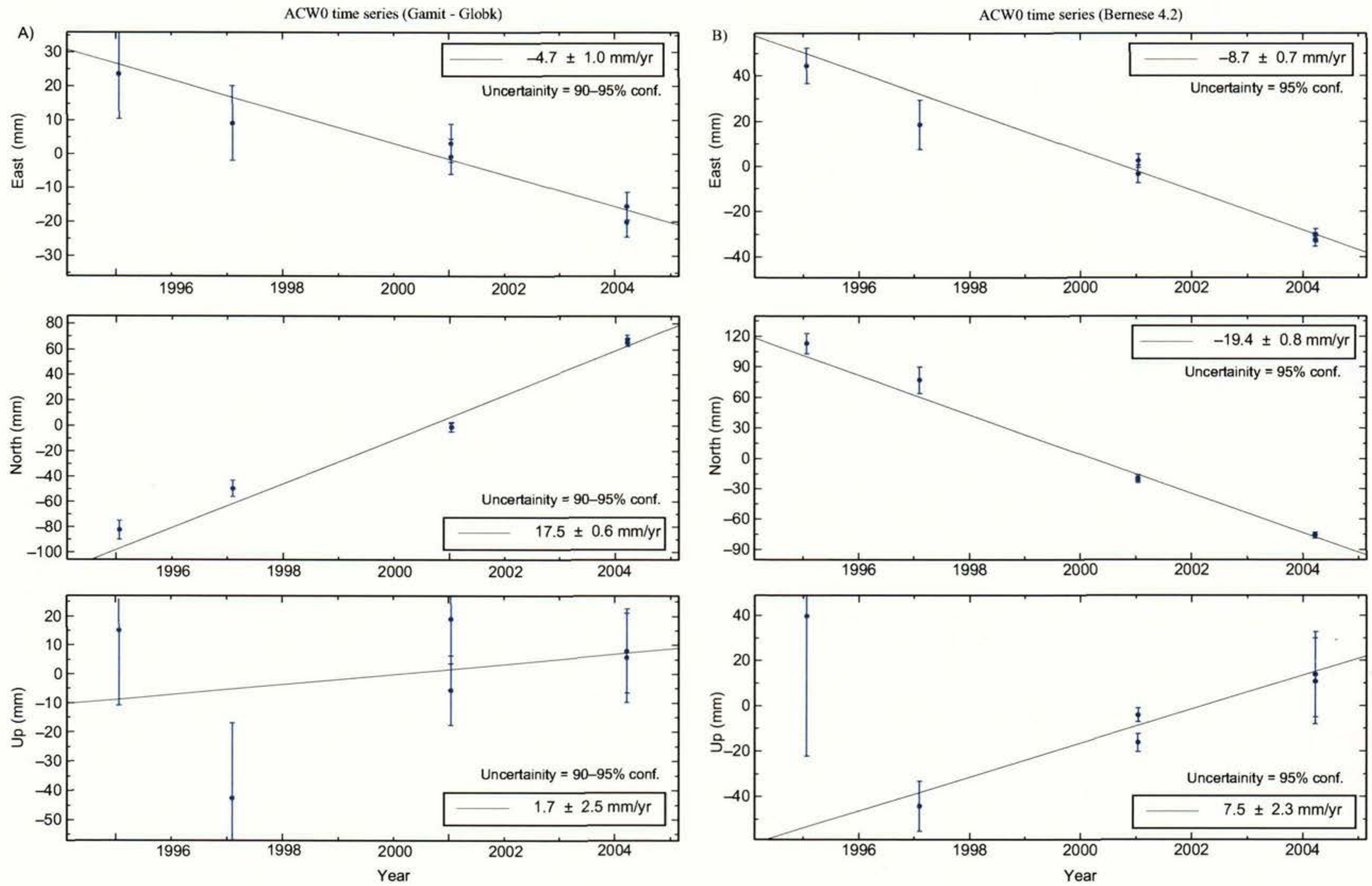

Figure 2.4: **ACW0 station position time series** A) GAMIT-GLOBK processing: station positions are relative to the ITRF2000 reference frame and uncertainties are 90%–95%. (GAMIT-GLOBK calculated uncertainties of 1σ were not scaled to account for $\chi^2 \approx 0.5$, thus uncertainties are approximately 90%–95%.) **B)** BERNESE-ADJCOORD processing: station positions are relative to AUCK and uncertainties are 95%.

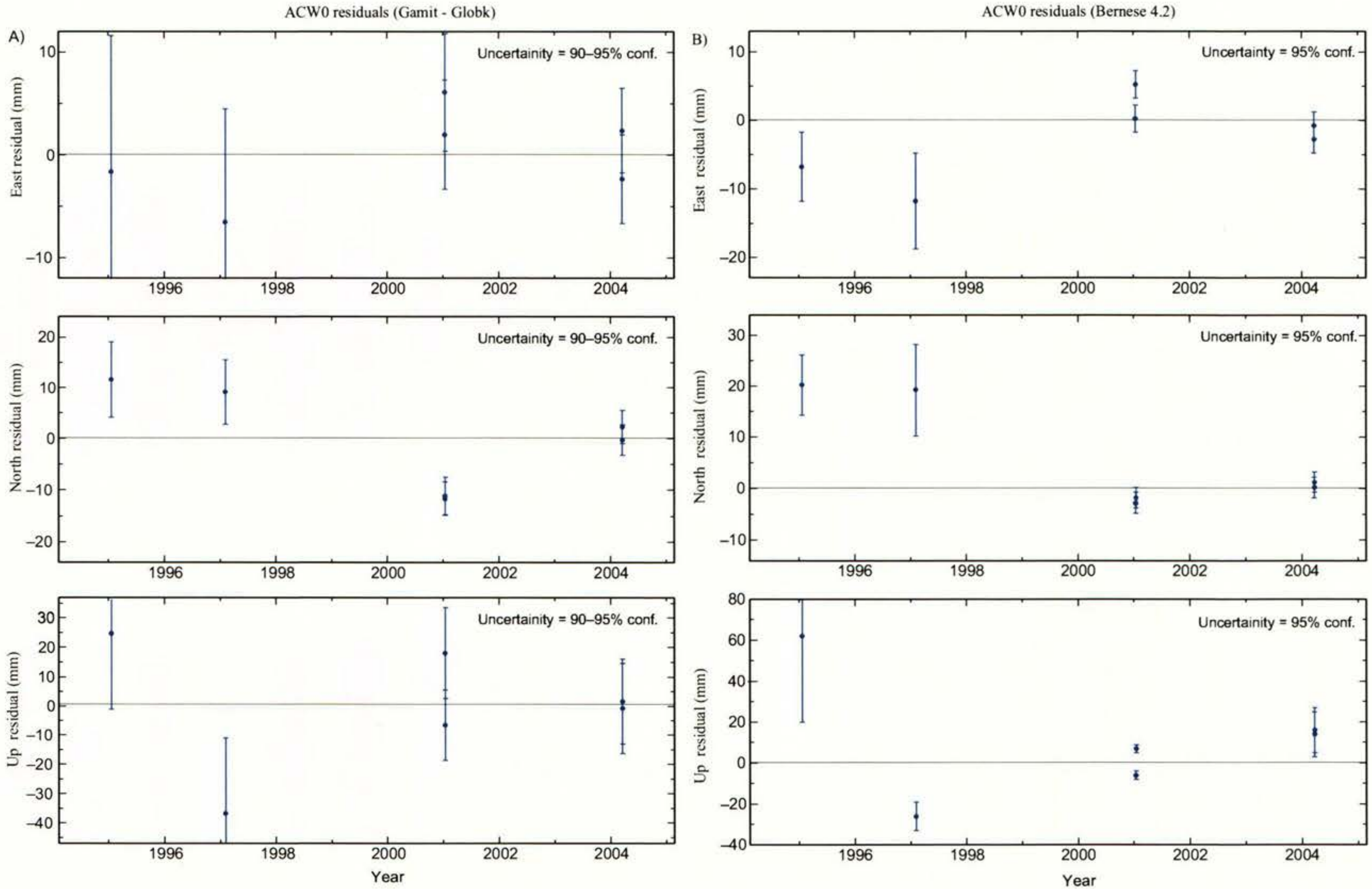

Figure 2.5: **ACW0 time series residuals** A) GAMIT-GLOBK processing: uncertainties are 90%–95%. (GAMIT-GLOBK calculated uncertainties of 1σ were not scaled to account for $\chi^2 \approx 0.5$, thus uncertainties are approximately 90%–95%.) **B)** BERNESE-ADJCOORD processing: uncertainties are 95%.

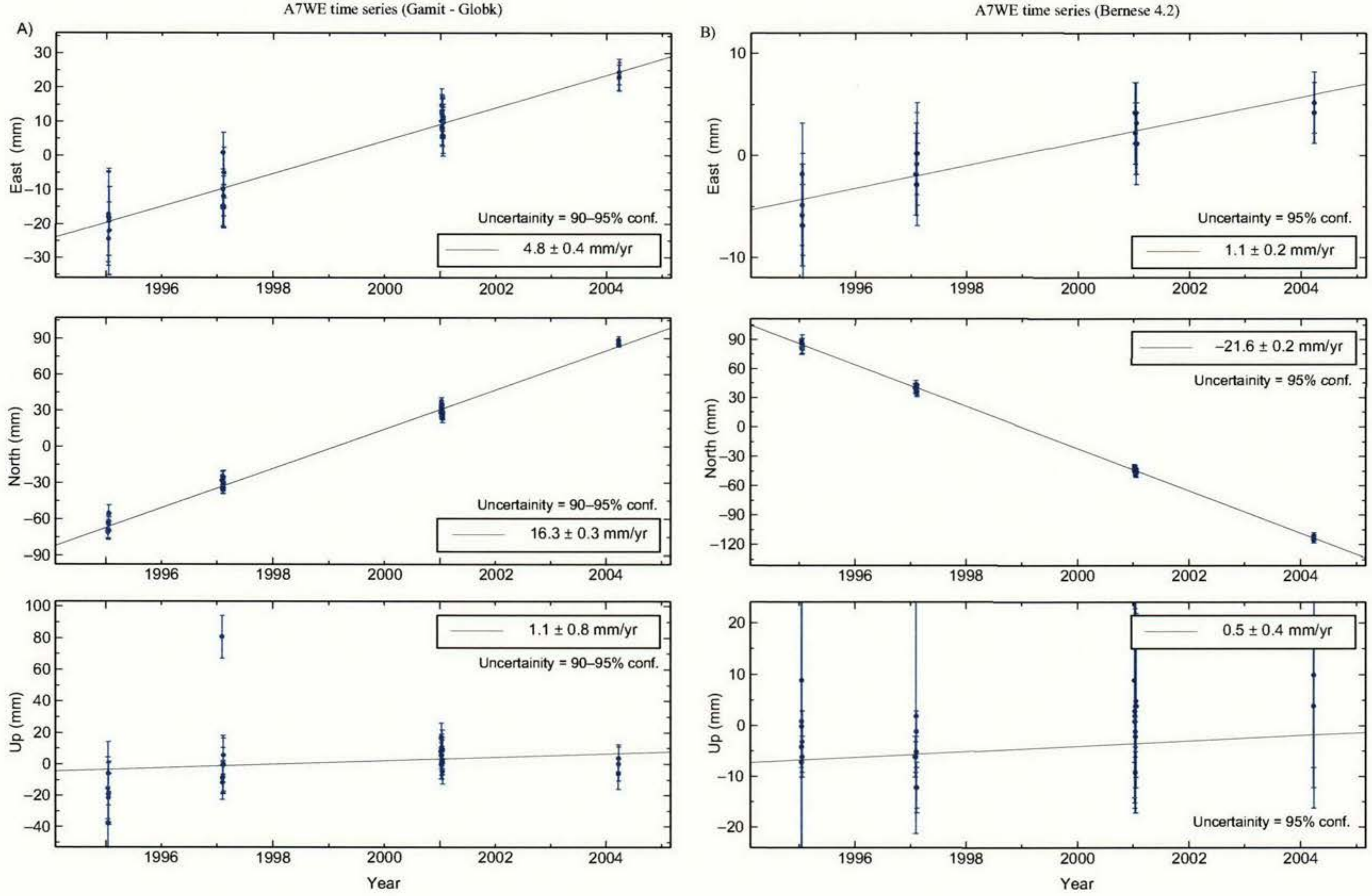

Figure 2.6: **A7WE station position time series** A) GAMIT-GLOBK processing: station positions are relative to the ITRF2000 reference frame and uncertainties are 90%–95%. (GAMIT-GLOBK calculated uncertainties of 1σ were not scaled to account for $\chi^2 \approx 0.5$, thus uncertainties are approximately 90%–95%.) B) BERNESE-ADJCOORD processing: station positions are relative to AUCK and uncertainties are 95%.

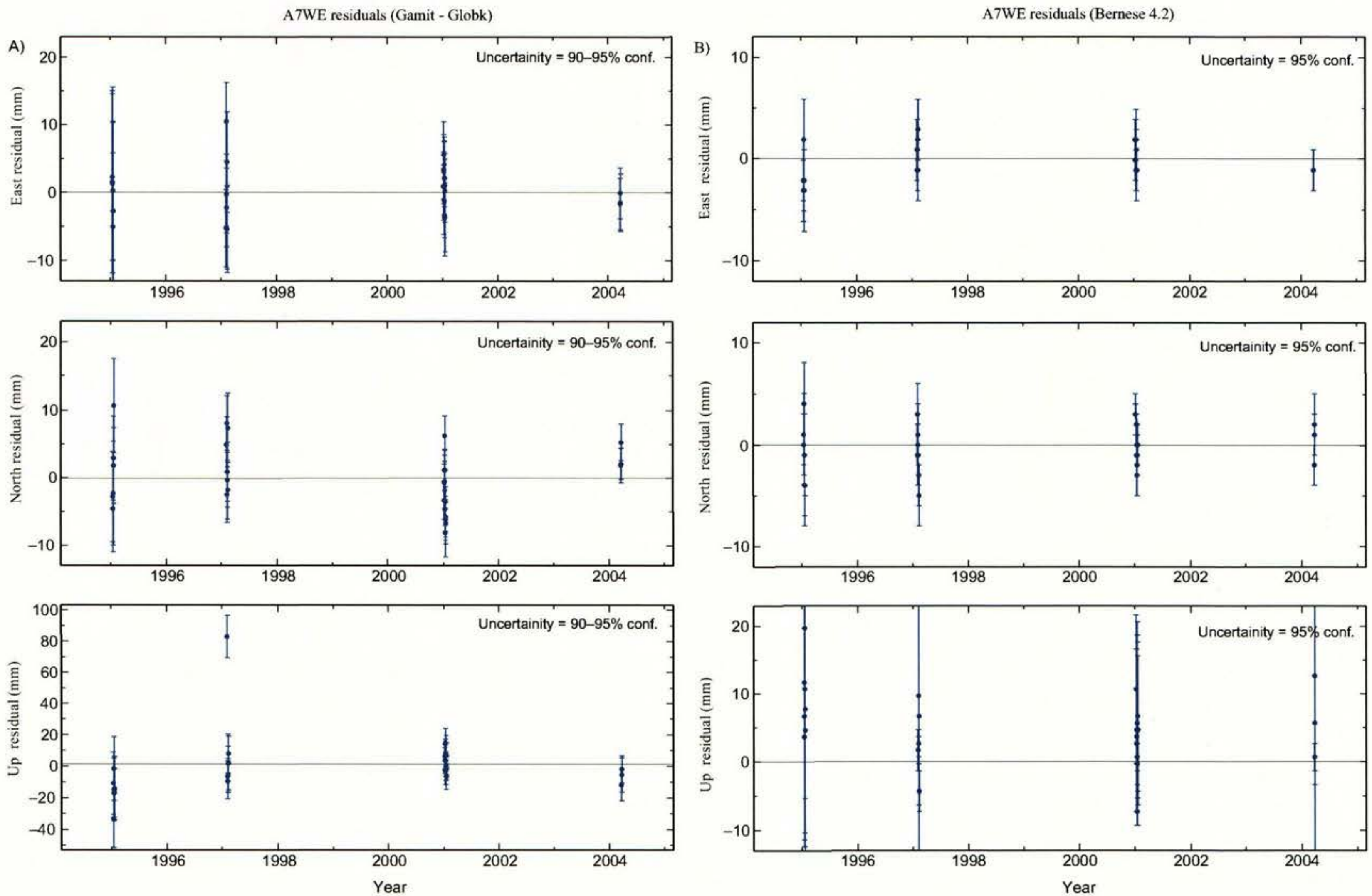

Figure 2.7: **A7WE time series residuals** A) GAMIT-GLOBK processing, uncertainties are 90%–95%. (GAMIT-GLOBK calculated uncertainties of 1σ were not scaled to account for $\chi^2 \approx 0.5$, thus uncertainties are approximately 90%–95%.) B) BERNESE-ADJCOORD processing: uncertainties are 95%.

BERNESE

Prior to BERNESE processing we constructed baselines between stations for each day of the campaign. The baselines specify which combinations of stations will be single-differenced in the processing sequence. For all campaign sites baselines were constructed to minimise the baseline length and maximise the observation time at each station. This was particulary important for the 1995 and 1997 campaign data when some stations were only observed for a very short time. Optimally, stations with poorer data quality are only used in one baseline.

Good a priori coordinates of at least one reference site must be known for successful processing (Beutler et al., 2001). The nearest continuous GPS station to Gisborne, which has been operating since 1995, is AUCK ($-36.60°$,174.83°, Figure 1.1), near Auckland. For processing, AUCK was constrained to a value within 2 mm of its ITRF coordinate. As AUCK was not operating in January 1995, the nearby station 5515 was tied to AUCK and used to calculate a 1995 location for AUCK.

Like GAMIT, BERNESE 4.2 comprises a series of modules which are run in sequence (Figure 2.3). In addition to the baseline and fixed reference site files, BERNESE 4.2 requires similar primary input files to those described for GAMIT.

BERNESE 4.2 can be divided into two parts — pre-processing and network solution estimation. Pre-processing programmes check and prepare the data for the main estimation programme (Beutler et al., 2001). Using the baseline files, IGS precise orbits, and constrained station coordinates in ITRF2000 as starting files we ran the standard BERNESE 4.2 processing using 30 s data through to the end of the single differencing and data cleaning phases. There are three main steps:

1. CODSPP—satellite clock errors are estimated using least squares adjustment and station positions estimated from pseudorange data;

2. SNGDIF—creates single difference observation files, forming baselines for double differencing;

3. MAUPRP—cycle slips are located and repaired, ensuring a continuous data stream, and outlier data points are rejected.

At this point we extracted coordinates for new stations from intermediary station coordinate files generated in the processing, and updated the starting coordinate files. The ionospheric delay is eliminated by processing double differenced L_3 phase observables.

In the main stage of processing the data are down-sampled to 120 seconds and a network solution is generated. The main processing occurs in parameter estimation programme GPSEST. First, an ionosphere-free solution is formed using the L_3 linear combination of L_1 and L_2 phase data. Station coordinates, tropospheric delay and phase ambiguities are all solved for. Next, double differenced phase ambiguities are resolved for each baseline using the wide-lane/narrow-lane technique, first estimating the wide lane value (L_5), then L_1 only (narrow lane). During the L_5 step, site coordinates are held fixed to their optimal values and an ionospheric model is used. In the narrow lane ambiguity resolution GPSEST is run to solve for L_1 ambiguities and the coordinates at the far end of each baseline. Finally, an ambiguity-fixed network solution is generated, using estimates from the previous steps, to solve for station coordinates and tropospheric delay.

ADJCOORD

ADJCOORD is a programme for the adjustment of geodetic survey coordinates to best fit survey data. A standard least-squares procedure is used to fit the coordinates and strain parameters to the survey data (Bibby, 1982). ADJCOORD determines the 95% confidence limits on standardised residuals of accepted observations and observations can be rejected if their standardised residual lies outside the 95% confidence interval (Bibby, 1982).

The output from the BERNESE 4.2 GPS processing is a set of daily files of coordinates and their associated covariances for each campaign. These are input into ADJCOORD where data outliers, which fall outside of the 95% confidence limits, are identified and may be selectively removed from processing. As we were looking for anomalous transient slip in the station position time series, we only removed coordinates from sessions with very short observation times (4–5 hours) or site velocities that were grossly inconsistent with neighbouring sites.

The ADJCOORD daily solutions for each campaign epoch were combined and we generated station position time series for individual campaign sites relative to AUCK (Figures 2.4B and 2.6B). Later we used the ADJCOORD output to generate displacement and velocity solutions for the campaign data set and strain rate solutions for regions of the Raukumara Peninsula (Chapter 4).

2.5 GPS processing outputs

The primary goal of our data analysis was to generate a coherent set of station position time series for the 1995–2004 period, which we could study for evidence of the non linear motion observed on continuous GPS during the Gisborne 2002 slow slip event (Chapter 3), and to look for evidence of other slow slip events prior to 2002. Selected station position time series resulting from the GAMIT-GLOBK and the BERNESE-ADJCOORD processing methods are shown in Figures 2.4 and 2.6. The GAMIT-GLOBK time series are relative to ITRF2000, while those from the BERNESE-ADJCOORD processing are relative to AUCK. Plotting the residuals of the station position time series for both the GAMIT-GLOBK and BERNESE solutions (Figures 2.5 and 2.7) illustrates that the results are reasonably consistent between the two sets of software. Plots of the residuals, calculated by removing the linear trend of the data, also help to identify non-linear motion in the time series as can be seen for site ACW0 in Figure 2.5. Processed campaign data are also used to examine spatial and temporal patterns in regional deformation over the Raukumara Peninsula (Chapter 4).

In general, the processing of local and regional New Zealand GPS data has been done using BERNESE software (e.g. Árnadóttir et al., 1999; Beavan and Haines, 2002). In order that our final published data set is compatible with, and comparable to, the larger New Zealand data set, we present analyses based on solutions from the BERNESE-ADJCOORD processing method in the rest of this thesis. Station position time series for Raukumara Peninsula campaign sites surveyed more than twice are presented in Appendix B.

Chapter 3

Gisborne 2002 slow slip event

We know from continuous GPS data that nonlinear deformation in the form of slow slip occurred in the Gisborne region in October 2002. In this chapter we examine the station position time series of campaign GPS data from the Raukumara Peninsula for evidence of the 2002 slow slip event and other events which may have occurred prior to 2002. Nonlinear slip should be recognisable in infrequently sampled measurements, such as those from the Raukumara Peninsula, as long as the data are represented as a station position time series rather than a velocity vector (e.g. Larson et al., 2004). We predict the recurrence interval for events with surface displacements similar in size to the Gisborne 2002 slow slip event and forward model the slip region of that event based on the continuous GPS data. Using continuous and campaign data, we investigate the effect of the Gisborne 2002 slow slip event on inter-seismic coupling and slip deficit rate on the northern Hikurangi subduction margin.

3.1 Time series

Station position time series (Figure 3.1) were generated using the results of the BERNESE-ADJCOORD processing, in which Auckland continuous GPS site AUCK was used as a base station. The time series show station locations relative to AUCK. Station positions at each time interval are plotted relative to the calculated mean station position and offset vertically for clarity. East, north and up components of all Raukumara Peninsula campaign sites surveyed more than twice are presented as time series in Appendix B.

Initially, in our close analysis of the campaign GPS time series, we assume that if a Raukumara campaign site has not been affected by a transient deformation such as a slow slip event its position will plot linearly in time. However, as Larson et al. (2004) discussed, it is very

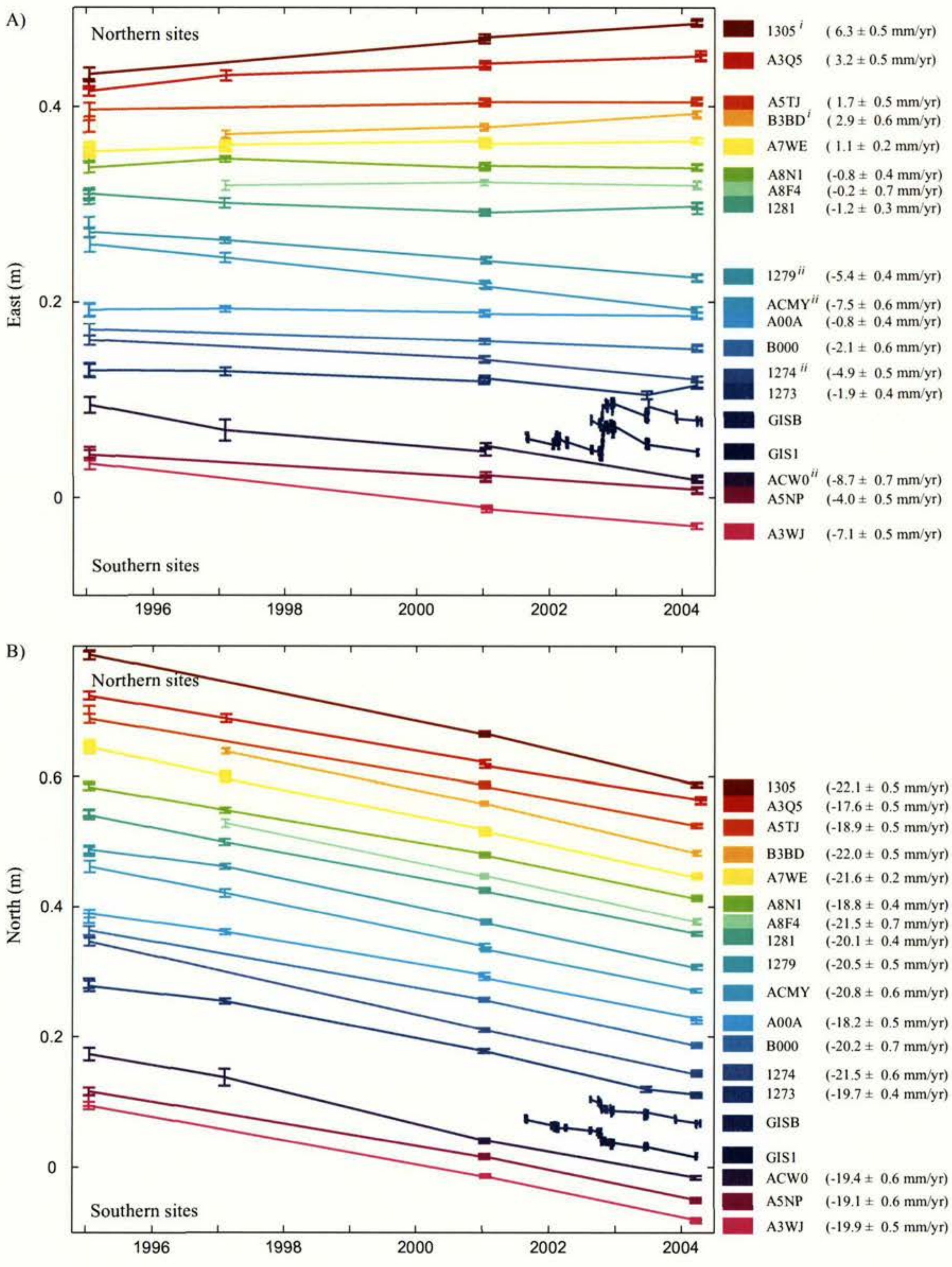

Figure 3.1: **Raukumara Peninsula GPS station position time series** for sites observed three times or more between 1995 and 2004, including selected days from continuous stations GIS1 and GISB. Time series are generated using data from the BERNESE-ADJCOORD processing, in which Auckland continuous GPS station, AUCK, was used as base station. We use the mean station position as the origin and plot deviations from that point. The time series are ordered south to north, with an arbitrary offset from the origin. Average rates of motion, relative to AUCK, are given for each station. They are calculated using a linear displacement model in ADJCOORD. A) East component; [i] East coast sites with rapid eastward displacements; [ii] East coast sites with rapid westward displacements (c.f. Figures 4.1– 4.4); B) North component.

important that the spatial and temporal sampling of the time series are adequate to distinguish steady-state from transient motions. Monument instability at the measurement site, or human error in campaign measurements may cause glitches in campaign time series, thus transient motions must also be larger than the uncertainties in the GPS positions to be recognised.

Figure 3.1 illustrates that the campaign GPS time series for the Raukumara Peninsula are approximately linear. We find that the long time intervals between GPS campaigns on the Raukumara Peninsula make it impossible to identify transient motion in the campaign time series. Even the 20 mm eastward displacement of the Gisborne 2002 slow slip event is not discernible in the campaign time series. Any eastward displacement caused by the slow slip event has been completely recovered by the rapid westward motion of the sites between events.

We therefore cannot use the campaign time series to help us constrain the spatial extent or moment release of the Gisborne 2002 slow slip event. Also, due to the low temporal spacing of observations in the campaign time series, slow slip events occurring previously could easily be overlooked. While there is no clear evidence for other slow slip displacements during the past ten years, changes in rate do occur in the campaign time series. In particular, between 1997 and 2001 there is an apparent slowing of westward motion and slight southward acceleration on site ACW0 (Figure 3.1 and 3.3), the closest site to GIS1. This apparent slowing of the westward rate and southward acceleration is consistent with a slow slip event of the same sense of displacement to the Gisborne 2002 event having occurred between 1997 and 2001.

3.2 Slow slip recurrence intervals

Quasi-periodic recurrence intervals have been observed for slow slip events in the Cascadia subduction margin (e.g. Miller et al., 2002; Szeliga et al., 2004), and the Guerrero region of Mexico (e.g. Lowry et al., 2005). Recurrence intervals have been used successfully in Cascadia to predict the timing of future events.

While we cannot use the campaign time series to effectively constrain the spatial extent of the Gisborne 2002 slow slip event, they do allow us to make estimates of the long term rates of motion for each campaign site on the Raukumara Peninsula. In turn, comparisons between the long term rates for campaign sites and short term inter-slip rates from continuous GPS stations allow us to make estimates for slow slip recurrence intervals. We use the term inter-slip to mean periods in the continuous GPS station position time series when no slow slip event is taking place and the observed station displacement is approximately linear (Figure 3.2). The inter-slip rate is estimated by averaging the best fit linear trends of the continuous GPS time series during those periods when no slow slip event is taking place.

One factor which may bias the recurrence interval calculations is the variation in inter-slip rates that occur in the continuous time series. Analysis of the GISB continuous time series following the Gisborne 2002 slow slip event show the westward rate of GISB to accelerate directly after the slow slip event (19.0$\pm$0.9 mm/yr west) until the smaller July 2003 slow slip event, after which the rate is slower (7.9$\pm$0.8 mm/yr west, Figure 3.2). No such acceleration was observed prior to the recent Gisborne 2004 event, rather the rate of motion for GISB appears to slow directly before the event (Figure 1.2).

In order to estimate a recurrence interval for slow slip similar in size to the Gisborne 2002 slow slip event we compare the long term rates from campaign time series spanning ten years with the short term inter-slip rates from the continuous GPS time series (Figure 3.3). We choose pairs of neighbouring stations ACW0 (campaign) and GIS1 ($\sim$5 km apart), and 1273 (campaign) and GISB ($\sim$10 km apart) (Figure 3.3), which are likely to have similar displacement histories. As stations 1273 and GIS1 are further apart we have less confidence in this comparison. Also, as we have only attained GPS data for site GIS1 at specific time intervals we use this decimated data instead of a full continuous time-series in our rate calculations.

We estimate an average inter-slip rate of 18.3$\pm$1.9 mm/yr to the west for continuous station GIS1 between the end of 2001 and March 2004 (Figure 3.3). Based on a long-term rate for site ACW0 of 8.7$\pm$0.7 mm/yr west between January 1995 and March 2004 we calculate that there is a 9.6$\pm$2.0 mm/yr deficit in westward motion in the ACW0 campaign data. In October 2002 the Gisborne slow slip event resulted in about 30 mm eastward displacement of GIS1. Based on these values the minimum recurrence interval for a similar slow slip would be 2.6 yrs, and the maximum, 3.9 yrs. The same calculations can be made comparing stations 1273 and GISB, although these stations are further apart than ACW0 and GIS1. The long term rate for station 1273 is 1.9$\pm$0.4 mm/yr west and the average inter-slip rate for GISB is 12.8$\pm$1.2 mm/yr west (Figure 3.3). There is a deficit in westward motion of 10.9$\pm$1.2 mm/yr between the two sites. The Gisborne 2002 slow slip event resulted in about 20 mm eastward displacement at GISB, thus we calculate a minimum recurrence interval here to be 1.7 yrs and a maximum of 2.1 yrs. The uncertainties in these recurrence estimates are probably underestimated as we have not incorporated the uncertainty in the magnitude of the slow slip events.

In general, our estimates point to slow slip events similar to the October 2002 event occurring every 2–4 yrs, which is in agreement with the most recent slow slip event which was

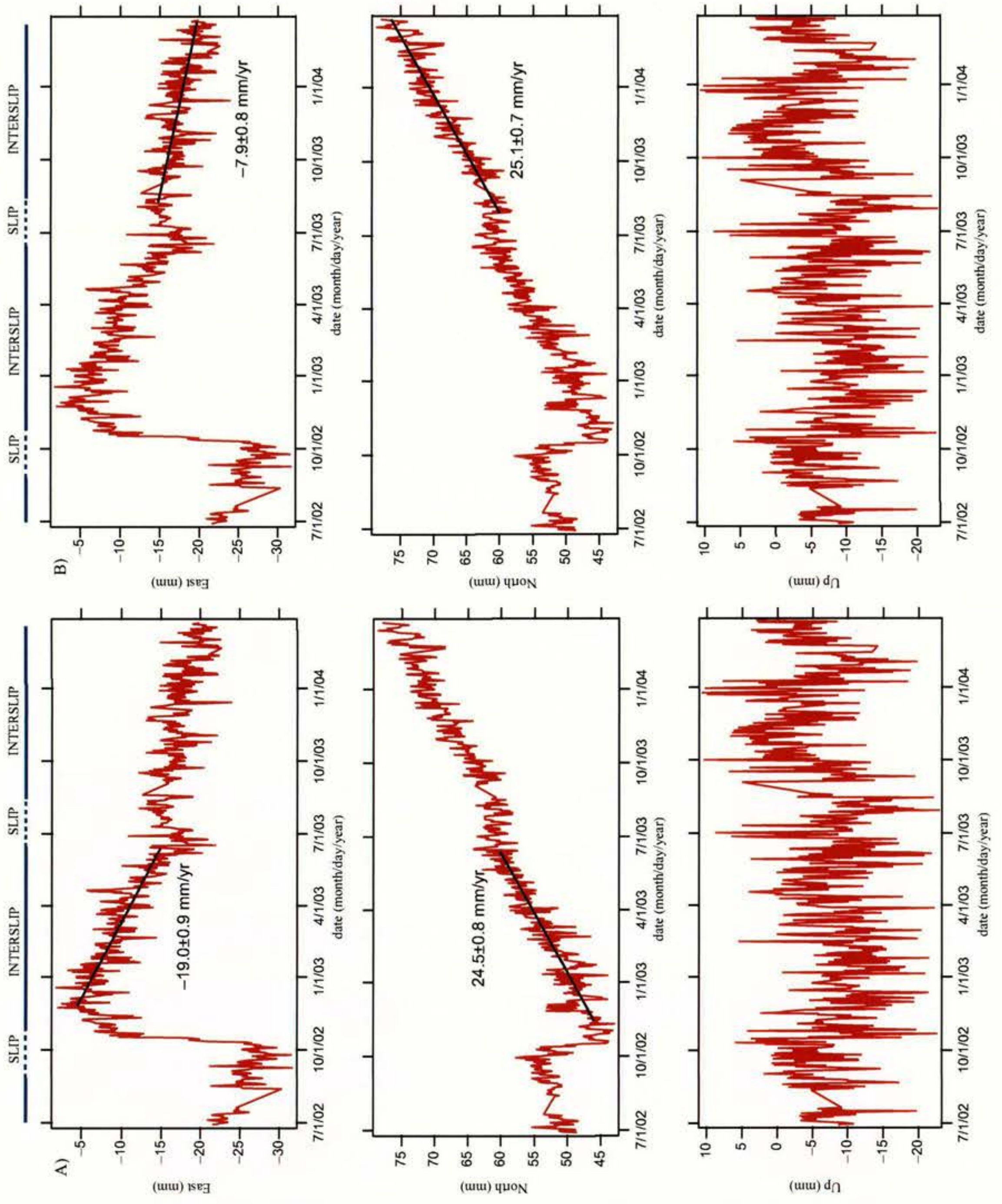

Figure 3.2: **Station position time series for continuous site GISB** showing variable rates of displacement, in particular accelerated westward rates of displacement directly before and after the Gisborne 2002 slow slip event. Slip and inter-slip periods are denoted by the dashed and solid lines respectively.

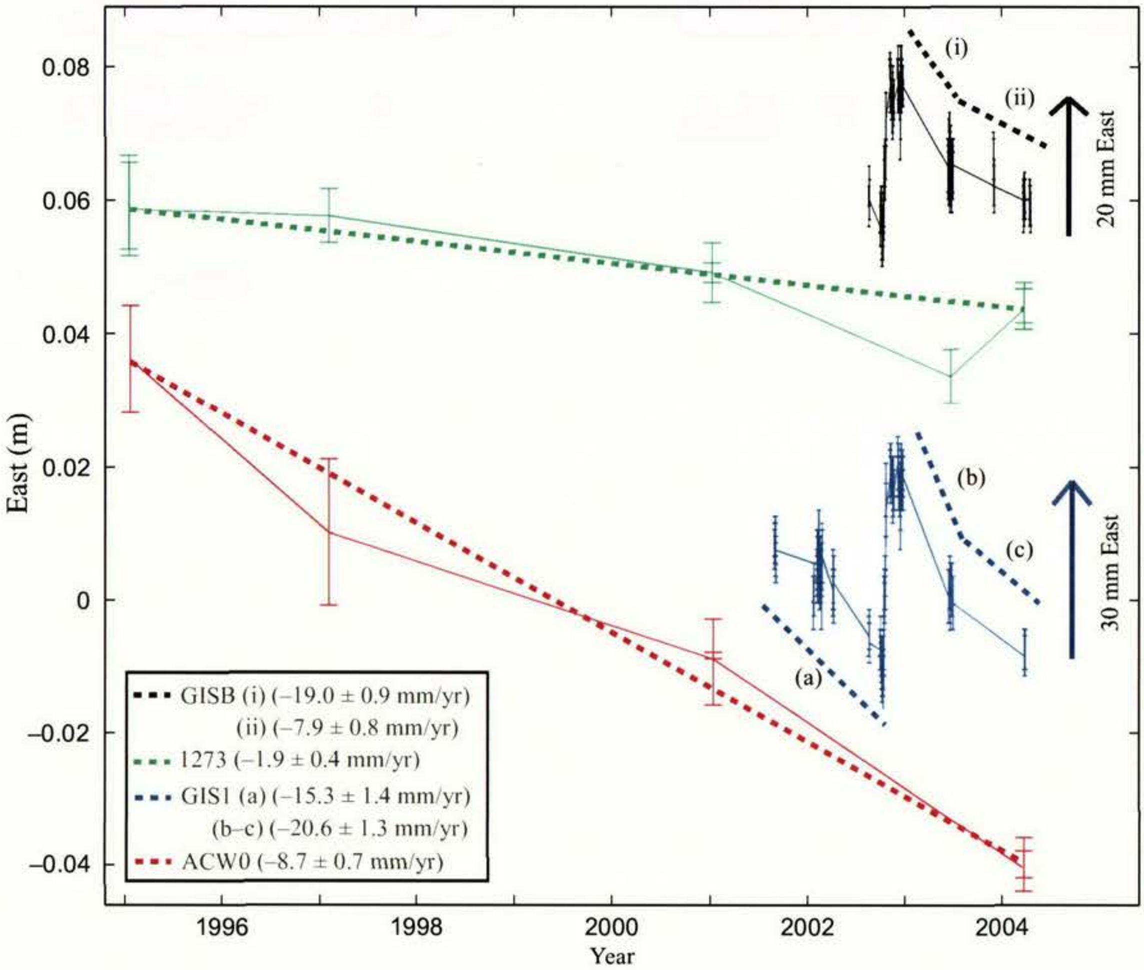

Figure 3.3: **Time series of neighbouring stations ACW0 and GIS1 (bottom), and 1273 and GISB (top)** used to calculate the recurrence interval for slow slip events of similar surface displacement to the Gisborne 2002 event. This comparison illustrates that such events could be easily overlooked in the campaign time series. Average long-term rates of motion are given for ACW0 and 1273; the rates given for GIS1 and GISB are best estimates of the inter-slip rate for each station; all rates are calculated using a linear displacement model in ADJCOORD. We use this information to calculate a recurrence interval for slow slip beneath the Raukumara Peninsula.

recorded in Gisborne in November 2004, just over two years after the previous event. Our tentative interpretation of a slow slip event occurring between 1997 and 2001, as implied from the apparent slowing in westward displacement in the ACW0 station position time series, would also be in accordance with this recurrence interval. We know from the continuous GPS record that smaller events, such as the 5 mm displacement recorded in July 2003, occur in the Gisborne region, thus the rate deficit seen on the campaign sites may be made up by a combination of large and small slow slip events.

3.3 Forward modelling of the Gisborne 2002 slow slip event

We apply a simple forward model to observed displacements from GPS continuous sites GIS1 and GISB from the Gisborne 2002 slow slip event (Figures 3.4 and 3.5). We are not able to constrain the model spatially using the campaign GPS time series; however, in Figure 3.4 we show model predictions at two campaign sites, 1273 and A8F4, to illustrate what effect the model slip could have over a larger region. We model slip on the subduction interface, which is a geometrically well defined structure through seismic studies (Reyners, 1998; Reyners and McGinty, 1999).

Our forward model uses nine preset parameters (Table 3.1), to predict the displacements that would be produced by slip on a rectangular region of the subduction interface. The depth and dip of the subduction interface were determined using subduction zone parameters from Reyners (1998). The azimuth of the slip plane is chosen to fit the subduction interface geometry and is not varied in the models. The dip, depth, latitude and longitude parameters are dependent on which part of the subduction interface the slip plane is located on. The depth parameter is the depth from the surface to the upper edge of the slip plane. We vary these four parameters (dip, depth, latitude and longitude) in accordance with the geometry for the subduction interface to allow us to model slip at different depths on the interface. The rake, length, width and slip parameters are varied through trial and error for each different slip plane location. We do not have any constraints from the geodetic data on the north-south extent of the slip plane.

Model A (Figures 3.4A and 3.5A), is a plane on the subduction interface west of Gisborne, which has an upper edge depth of 19 km. Thrust slip of 30 cm on this plane (20 km $\times$ 40 km) results in a good fit to the horizontal data, but predicts an upwards vertical displacement for the continuous sites that was not observed. Model B (Figures 3.4B and 3.5B), with an upper edge depth of 9 km, lies on the subduction interface offshore of Gisborne. Thrust slip of 18 cm on this plane (25 km $\times$ 60 km) fits the observed horizontal data quite well, and also agrees with

Model	Azimuth	Dip	Rake	Length (L) (km)	Width (w) (km)	Slip ($\bar{D}$) (m)	Depth (km)	Latitude	Longitude
A	205	14	90	40	20	0.3	19	–38.6	177.9
B	205	10.5	100	60	25	0.18	9	–38.9	178.4

Table 3.1: **Model parameters for slow slip on the subduction interface** of the Hikurangi subduction margin. The subduction zone parameters are taken from Reyners (1998) and Reyners and McGinty (1999).

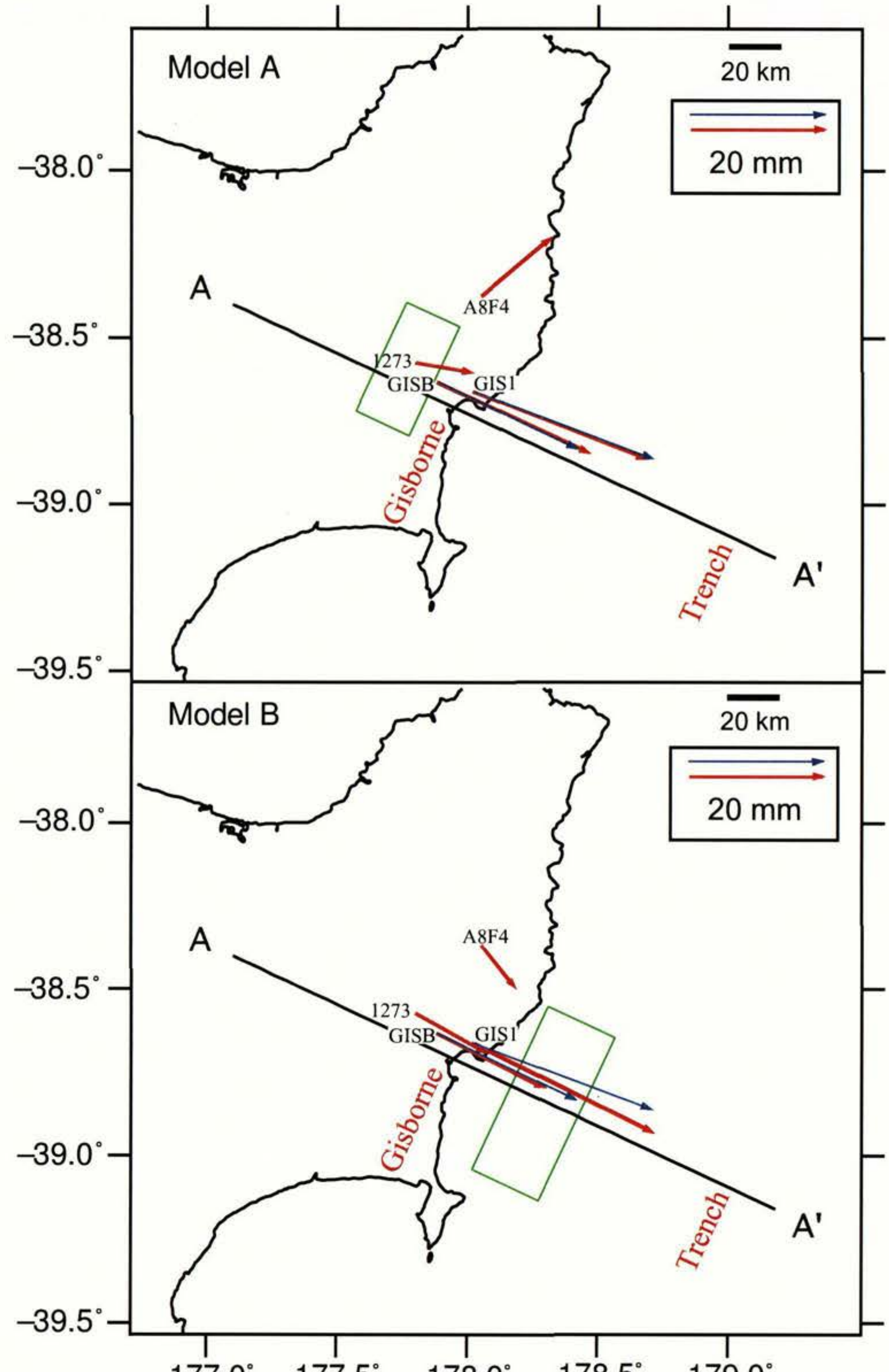

Figure 3.4: **Slip planes for two different forward models of the Gisborne 2002 slow slip event**. Both model slip planes are shown in green and lie on the subduction interface; blue vectors are the observed station displacements, red vectors are the station displacements predicted by the model. Model A) Deep subduction interface model; Model B) Shallow subduction interface model. Models are based on observations from continuous GPS sites GIS1 and GISB only. Profiles along cross section A–A' are shown in Figure 3.5. Model B is our preferred model.

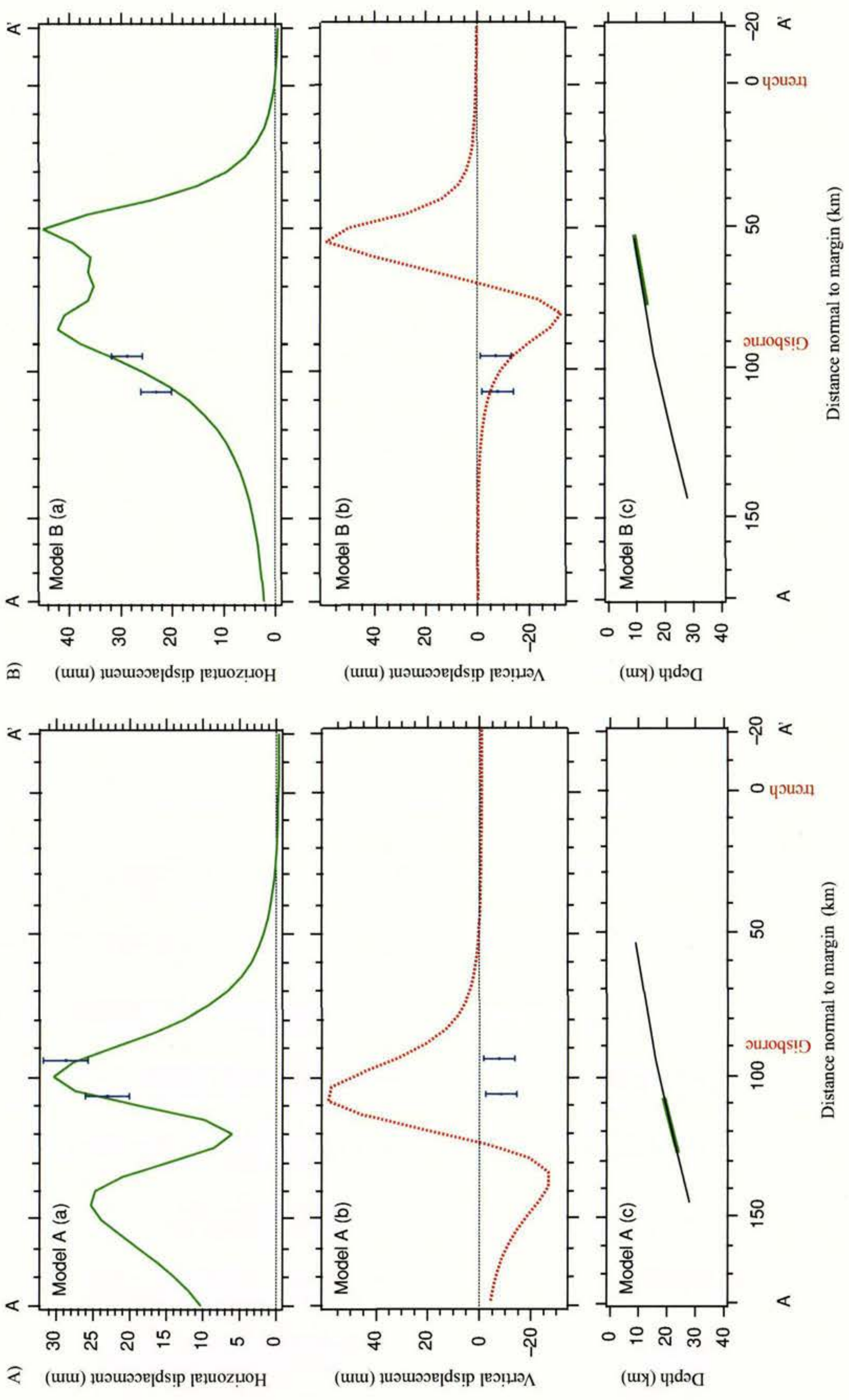

Figure 3.5: **Models A and B: Profiles of observed and predicted displacements (mm)**; (a) Horizontal displacement model with observed data points; (b) Vertical displacement model with observed data points; (c) Location of slip plane on the subduction interface. The map location of cross section A–A' is shown in Figure 3.4. Model B is our preferred model.

the downward displacement of the Gisborne continuous GPS sites. We favour Model B and rule out the deeper subduction model (Model A), as it predicts uplift for the Gisborne region where downwards displacements were observed. The moment released by a slow slip event with the slip parameters of Model B would be 0.81×10^{19} N m.

While fault planes smaller than that in (Figures 3.4B) can fit the data equally well, the larger slip required, combined with our estimated slow slip recurrence interval of 2–3 yrs, implies that the rate of slip in slow events would exceed the long term rate of convergence between the Pacific plate and the forearc block of ~54 mm/yr up-dip (Wallace et al., 2004).

The slip region in our preferred model, Model B, is towards the lower end of the seismogenic zone on the subduction interface (Figure 3.6). In comparison to events recorded on the Cascadia margin, Guerrero Gap or Japan (e.g. Dragert et al., 2001; Obara et al., 2004; Larson et al., 2004), our modelling suggests that the Gisborne slow slip events are occurring at shallower depths on the subduction interface. In all cases however, slow slip occurs below the main seismogenic zone on the plate interface, it is just that the downdip end of inter-seismic coupling on the north-eastern Hikurangi subduction zone is comparatively shallow (Reyners, 1998; Wallace et al., 2004).

3.4 Inter-seismic coupling and slip deficit rate

Modelling by Wallace et al. (2004) of inter-seismic coupling along the Hikurangi Margin shows that the Gisborne area straddles a small patch of elevated interseismic coupling (Figure 3.6).

Based on geodetic, geologic, and seismic data, Wallace et al. (2004) divided the eastern North Island into tectonic blocks. They implemented an elastic, rotating block approach, (McCaffrey, 1995, 2002), to invert for the angular velocities of each block, as well as the degree of coupling on faults bounding the blocks, including the Hikurangi subduction zone. In this project we are interested in the time-variance of inter-seismic coupling and slip deficit rate models along the Hikurangi margin, and for the Raukumara region in particular. We solve for inter-seismic coupling on the Hikurangi subduction zone, following the model of Wallace et al. (2004). For further details of the modelling approach and implications for the interpretation of North Island regional tectonics we refer the reader to Wallace et al. (2004). The degree of coupling (Phi) is specified by a number from zero (no coupling) to one (full coupling or locked) (Figure 3.6). The slip deficit rate is the rate of relative block motion that is not accommodated by aseismic creep in the inter-seismic period (Figure 3.7, Wallace et al., 2004).

In order to investigate the effect of the Gisborne 2002 slow slip event on the inter-seismic coupling and slip deficit rate for the Raukumara Peninsula region we run three different iterations of the Wallace et al. (2004) model. The first iteration (Figures 3.6A and 3.7A) is the same as the published model from Wallace et al. (2004), using GPS campaign data up to 2001. This model shows the inter-seismic coupling and slip deficit on the Hikurangi margin prior the Gisborne slow slip event. The second iteration (Figures 3.6B and 3.7B) models data from the 2001–2004 period only, to see if there was a marked change in coupling compared to the 1995–2001 period. This iteration of the model (Figure 3.6B and 3.7B) shows the state of inter-seismic coupling and slip deficit on the Hikurangi margin between 2001 and 2004. Thirdly, we model all available campaign GPS data (1995–2004) for the Raukumara Peninsula (Figure 3.6C and 3.7C) to show the average inter-seismic coupling and slip deficit for the whole period.

We find that there are some differences between the three separate iterations of the slip deficit rate models. In all cases there is a zone of elevated inter-seismic coupling and elevated slip deficit on the Hikurangi subduction interface beneath the broader Gisborne region. For the 1995–2001 model iteration (Figures 3.6A and 3.7A) the zone of elevated inter-seismic coupling and slip deficit forms a broad diffuse region under Gisborne and there is a strong band of higher slip deficit to the north, perpendicular to the coast under Tolaga Bay. In this model there is a gradual increase in inter-seismic coupling and slip deficit toward the trench. In the 2001–2004 model iteration (Figures 3.6B and 3.7B) there is no zone of elevated inter-seismic coupling and slip deficit under the Gisborne region. The slip deficit increases abruptly at the eastern edge of the modelled slip plane, to a higher level than for the 1995—2001 model, and there is no strong band of of elevated slip deficit under the Tolaga Bay area. The third model, which includes data from all years, is an average of the two patterns.

The apparent variation between the three model iterations may be due to lower spatial density of the data in the 2001–2004 model. Moreover, there are no data points offshore and the resolution of the model decreases away from the coast; and there may be a trade-off in the model between high coupling offshore and lower coupling onshore, which the model distributes differently between epochs. In addition, the broad regional scale of the model may not resolve the local variations in deformation very accurately. For these reasons we must be cautious with our interpretation of the separate model iterations.

In general, our models show that a narrow zone of elevated coupling and slip deficit beneath

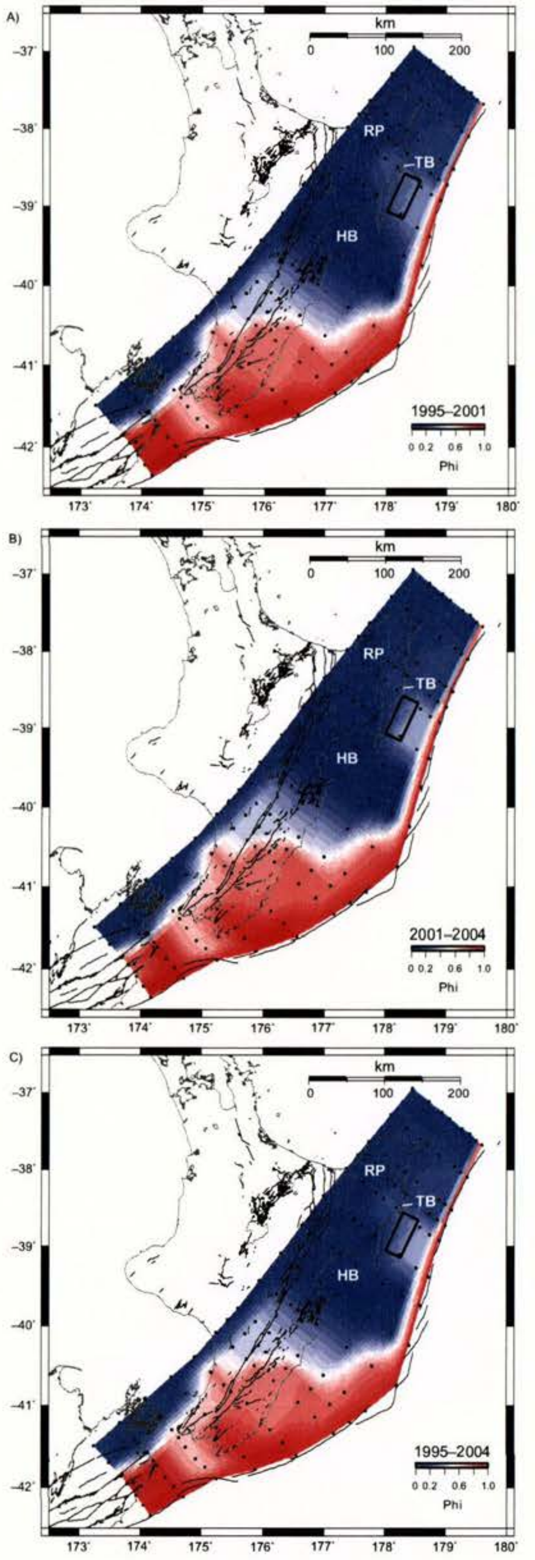

Figure 3.6: **Inter-seismic coupling (phi) on the Hikurangi subduction margin**, after Wallace et al. (2004). A) 1995–2001; B) 2001–2004; C) 1995–2004. The slip plane for Model B is shown as a black rectangle; it coincides with the downdip end of inter-seismic coupling on the subduction interface. Individual nodes are specified on the subduction zone and the degree of coupling at each node is solved for (Wallace et al., 2004). RP - Raukumara Peninsula; HB - Hawke Bay; TB - Tolaga Bay.

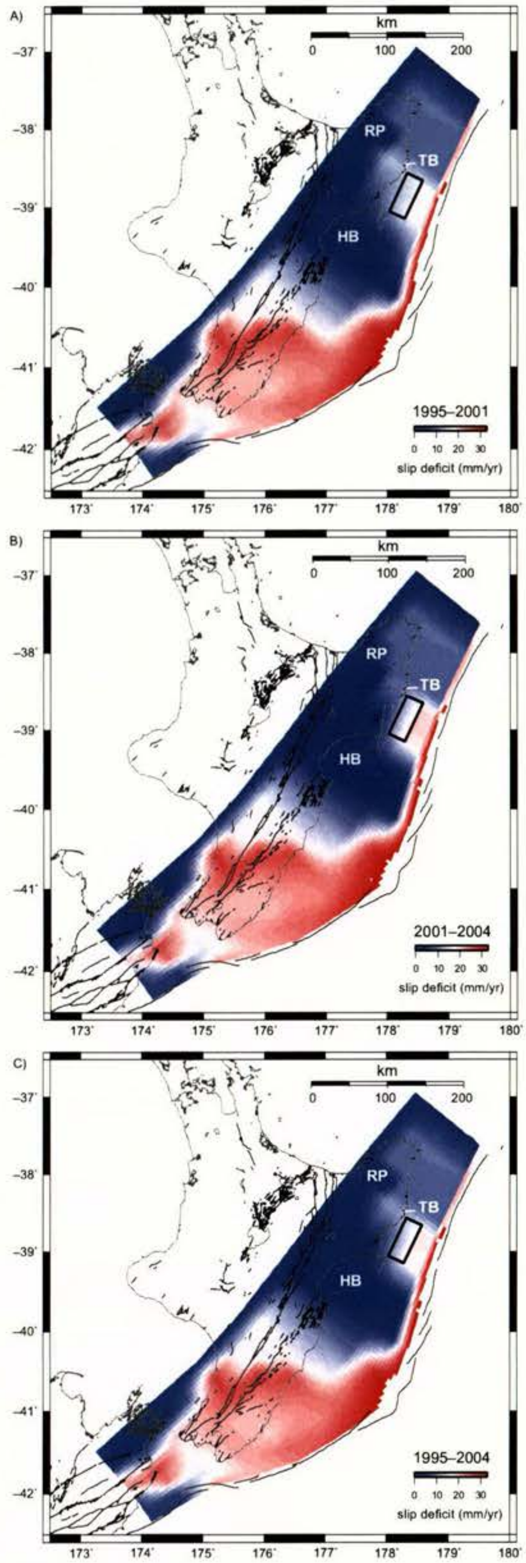

Figure 3.7: **Slip deficit rate mm/yr for Hikurangi subduction margin**, after Wallace et al. (2004). A) 1995–2001; B) 2001–2004; C) 1995–2004. The slip plane for Model B is shown as a black rectangle; it coincides with the downdip end of a zone of elevated slip deficit on the subduction interface. RP - Raukumara Peninsula; HB - Hawke Bay; TB - Tolaga Bay.

a coastal section of the Raukumara Peninsula, between Gisborne and Tolaga Bay, persists in spite of motion between the Australian and Pacific plates being intermittently accommodated through slow slip events on the subduction interface. The slip deficit may be somewhat reduced in places through slow slip, but not eliminated.

Slow slip events on the northern Hikurangi margin may push the zone of elevated coupling on the subduction interface updip and closer to failure through earthquake. Two tsunamigenic earthquakes, which occurred on the shallow outer subduction interface (Figure 1.4) in 1947 were interpreted to have been triggered by a slow slip event, which extended over the whole seismogenic zone of the subduction interface below the Raukumara Peninsula (Árnadóttir et al., 1999; Downes et al., 2004). There is evidence from around the Pacific rim for causal relationships between slow slip events and earthquakes. For example, an slow slip event on the Guerrero Gap in 2001, was followed by a $M_w = 6.7$ tsunamigenic earthquake located near the trench (Kostoglodov et al., 2003). No large earthquakes were associated with either the 2002 or 2004 Gisborne slow slip events.

3.5 Summary

The low temporal spacing of observations in the campaign station position time series aliases the Gisborne 2002 slow slip event, in which 20 mm surface displacement was recorded at continuous GPS station GISB. We cannot use the campaign GPS station position time series to constrain the spatial extent or moment release of this event. However, we do use them to make estimates of the recurrence interval for events of similar surface displacement to the 2002 slow slip occurring in the Gisborne region. Our calculations show that such events should occur at 2–3 year intervals, which is in agreement with the recent November 2004 slow slip event in the Gisborne region. We also tentatively interpret the slowing of westward motion in the ACW0 station position time series between 1997 and 2001 to be due to a slow slip event.

We model a slip plane for the Gisborne 2002 slow slip event using a simple forward model based on the continuous GPS data and the well known geometry of the subduction interface. Our favoured model is for 18 cm of slip on a plane, 60 km $\times$ 25 km, on the subduction interface offshore of Gisborne, though the fault length is poorly constrained. The model slip plane coincides with the downdip limits of the seismogenic zone on the subduction interface from Reyners (1998) and Wallace et al. (2004).

Models of inter-seismic coupling and slip deficit on the northern Hikurangi margin show a zone of elevated coupling beneath the Tolaga Bay area of the Raukumara Peninsula. This zone persists in spite of the relative plate motion between the Australian and Pacific plates being intermittently accommodated through slow slip events on the subduction interface. Two tsunamigenic earthquakes in 1947, which were located on the shallow subduction interface near the trench (Árnadóttir et al., 1999; Downes et al., 2004), may have been triggered through a slow slip event increasing stress on the upper part of the plate interface .

Chapter 4

Regional deformation

The campaign GPS record complements the short, two year history of the continuous GPS records, and allows more detailed spatial coverage of the Raukumara Peninsula than is currently practicable with continuous GPS. Displacement and velocity fields from the Raukumara Peninsula campaign GPS data provide us with a regional overview of spatial and temporal deformation patterns. We make comparisons, based on the most recently collected data, with a study of strain over the Raukumara Peninsula by Árnadóttir et al. (1999).

4.1 Displacement and velocity solutions

Total displacement (Figures 4.1 and 4.2) and velocity fields (Figures 4.3 and 4.4) for the 1995–2001 and 2001–2004 epochs of campaign GPS data for the Raukumara Peninsula show how deformation on the Raukumara Peninsula varies with space and time. The total displacement fields allow us to assess deformation occurring within subregions of the peninsula used for strain analysis (Figures 4.5 and 4.6).

We calculate the displacement solution using a two dimensional adjustment in ADJCOORD (Crook, 1992). We use campaign site coordinates calculated in the BERNESE processing and take into account the horizontal data components from specified epochs only. The solution assumes a planar surface, which works well for a small area such as the Raukumara Peninsula.

Velocity solutions are also computed in ADJCOORD, using a three dimensional adjustment and a specified time model, assuming constant station velocities (Crook, 1992). We choose to plot displacements and velocities relative to the average velocity of four western Raukumara Peninsula campaign GPS stations — 1297, A3Q5, A8N1, and A00A (Figure 4.1) — which

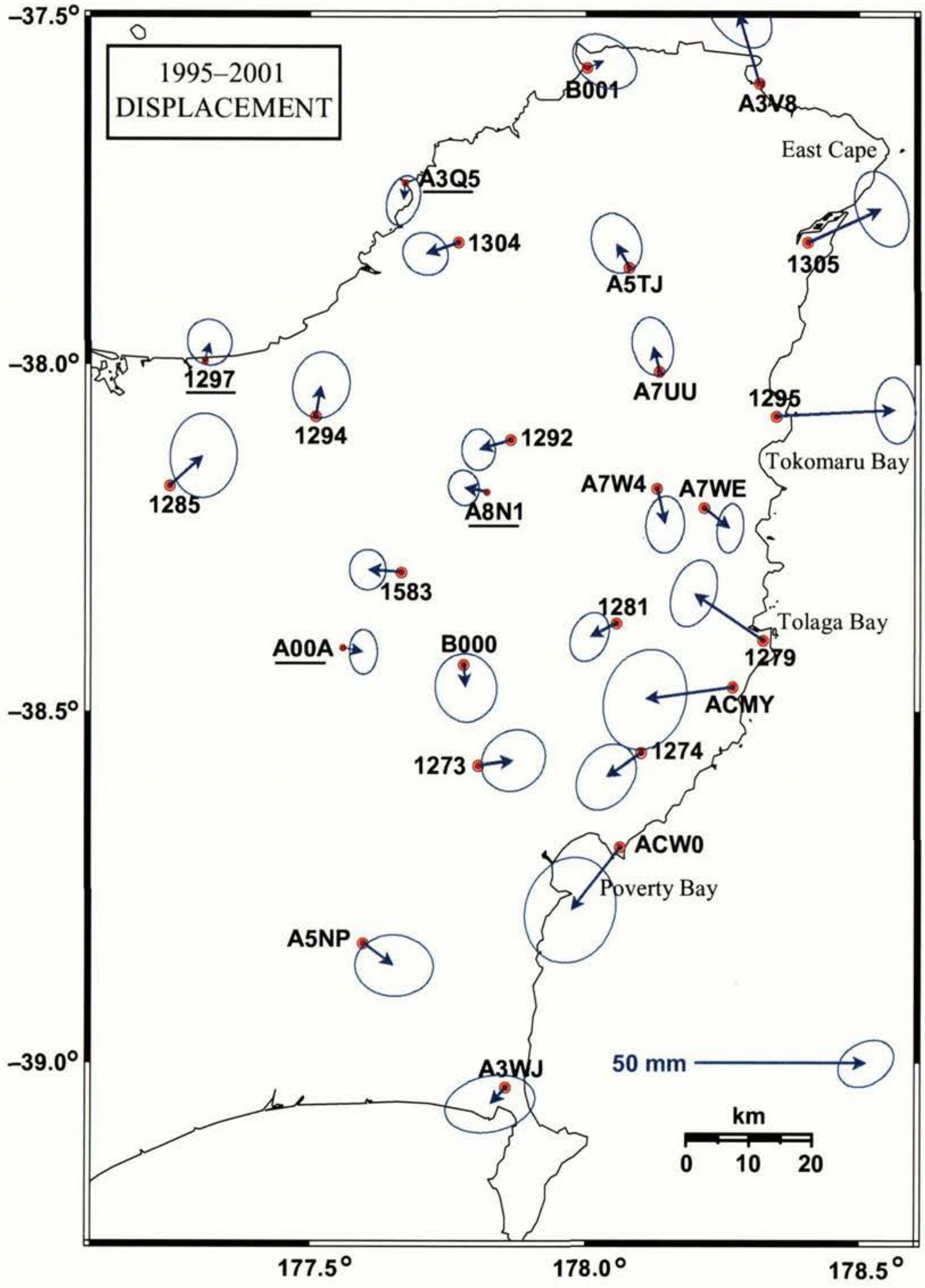

Figure 4.1: **Displacement solution for Raukumara Peninsula GPS sites between 1995 and 2001**, calculated in two dimensions. The displacement solution is plotted relative to western Raukumara Peninsula stations 1297, A3Q5, A8N1, and A00A (underlined). A rotation between surveys has been calculated. Rotation of the second survey coordinates (2001) = -0.4949 μrad. Uncertainty ellipses are 95% conf.

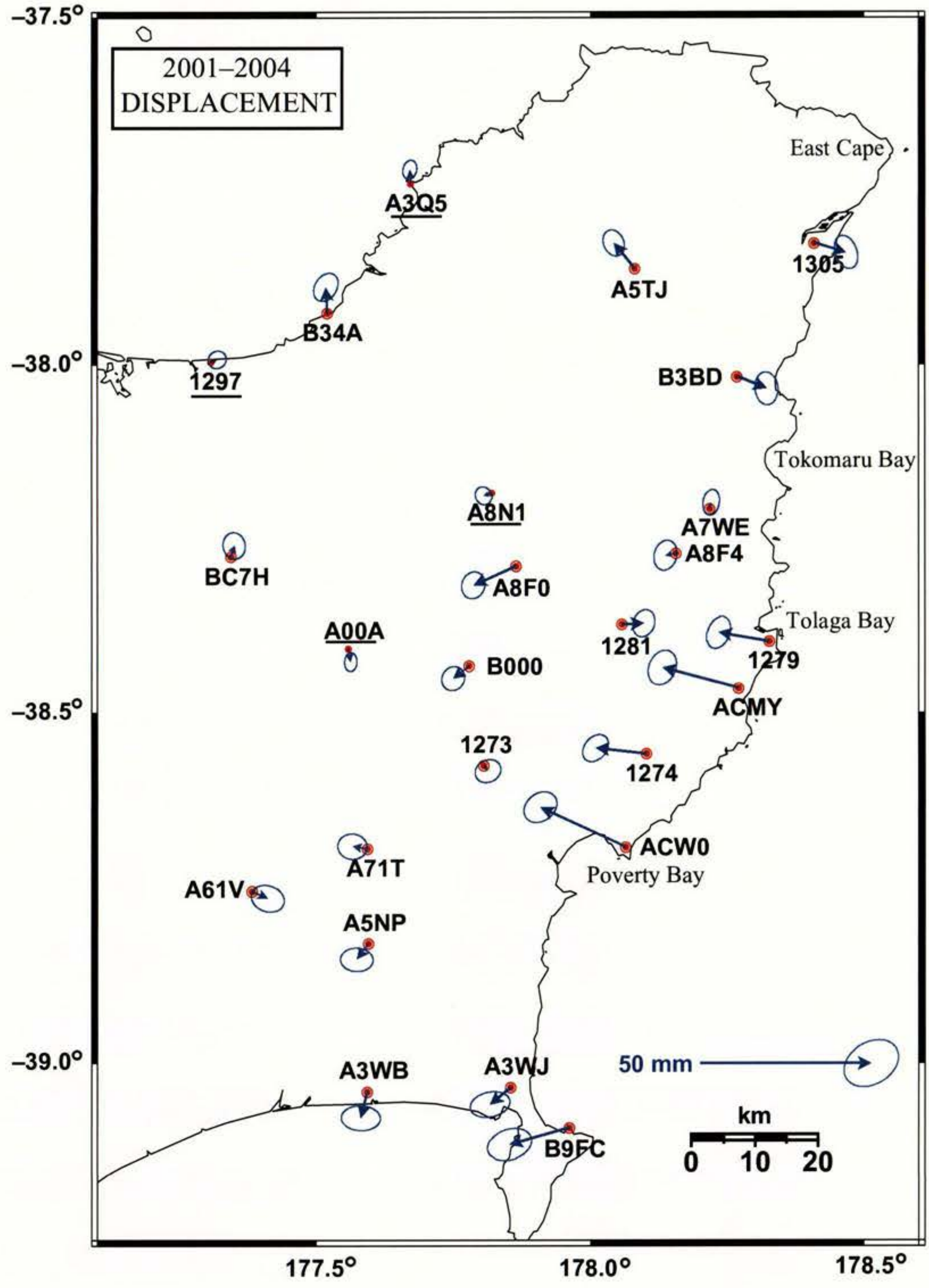

Figure 4.2: **Displacement solution for Raukumara Peninsula GPS sites between 2001 and 2004**, calculated in two dimensions. The displacement solution is plotted relative to western Raukumara Peninsula stations 1297, A3Q5, A8N1, and A00A (underlined). A rotation between surveys has been calculated. Rotation of the second survey coordinates (2004) = -0.1564 μrad. Note that this displacement solution shows the same information as the 2001–2004 velocity solution in Figure 4.4. Uncertainty ellipses are 95% conf.

are sufficiently removed from the Gisborne region that they were probably not affected by the 2002 slow slip event. The amount of rotation that has occurred between survey epochs is calculated in ADJCOORD and removed from the solution. Uncertainties for the 1995 and 1997 data sets are much greater than for either 2001 or 2004, so the 1995–1997 displacement and velocity solutions are not presented here.

The displacement and velocity solutions show three distinct regions of deformation on the Raukumara Peninsula (Figures 4.1– 4.4):

1. On the east coast between Poverty Bay and Tolaga Bay, campaign sites ACMY and 1279 show consistently high rates of displacement to the west–northwest. ACW0 and 1274 have high rates of displacement to the southwest during the 1995–2001 epoch and during the 2001–2004 epoch these same sites move consistently west–northwest parallel to ACMY and 1279. In general, all four sites have higher rates of westward displacement in the 2001–2004 epoch than in the 1995–2001 epoch.

2. On the east coast north of Tokomaru Bay, campaign sites 1295 and 1305 exhibit rapid eastward motion. Site 1295 was not observed in the 2004 campaign.

3. The central and western Raukumara Peninsula campaign sites have much smaller displacements than the east coast sites and show no clear pattern of deformation.

4.1.1 Interpretation of the displacement and velocity solutions

Displacement and velocity vectors show that the east coast section of the Raukumara Peninsula is rapidly deforming compared to the more stable central and western regions of the peninsula. We view this pattern of deformation in the context of inter-seismic coupling on the subduction interface beneath the Raukumara Peninsula.

The coastal sites between Poverty Bay and Tolaga Bay show rapid westward displacement. A narrow zone of higher inter-seismic coupling on the subduction interface below this region is interpreted to be the cause of this dominant westerly direction of motion, consistent with the westward subduction of the Pacific plate under the North Island (Figure 3.6, Wallace et al., 2004). This westward motion is punctuated by slow slip events, during which inter-seismic coupling is reduced, allowing the plates to slide past each other generating an eastward displacement of the Gisborne region as seen in the continuous GPS record.

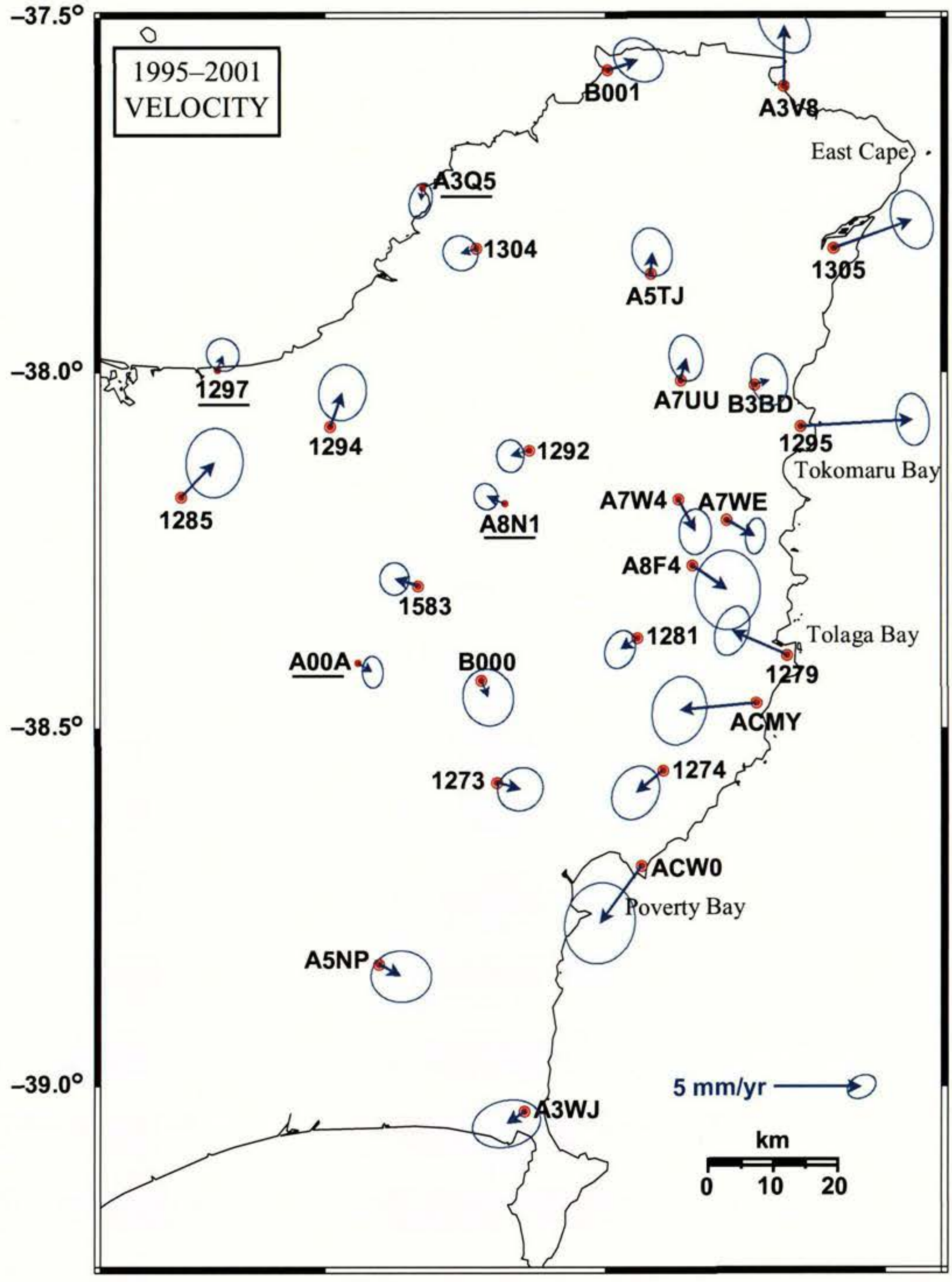

Figure 4.3: **Velocity solution for Raukumara Peninsula GPS sites between 1995 and 2001**. The solution includes site positions from the 1997 campaign. The velocity solution is plotted relative to western Raukumara Peninsula stations 1297, A3Q5, A8N1, and A00A (underlined). A rotation between surveys has been calculated and removed. Rotation of the second survey coordinates (2001) = -0.0749 μrad. Uncertainty ellipses are 95% conf.

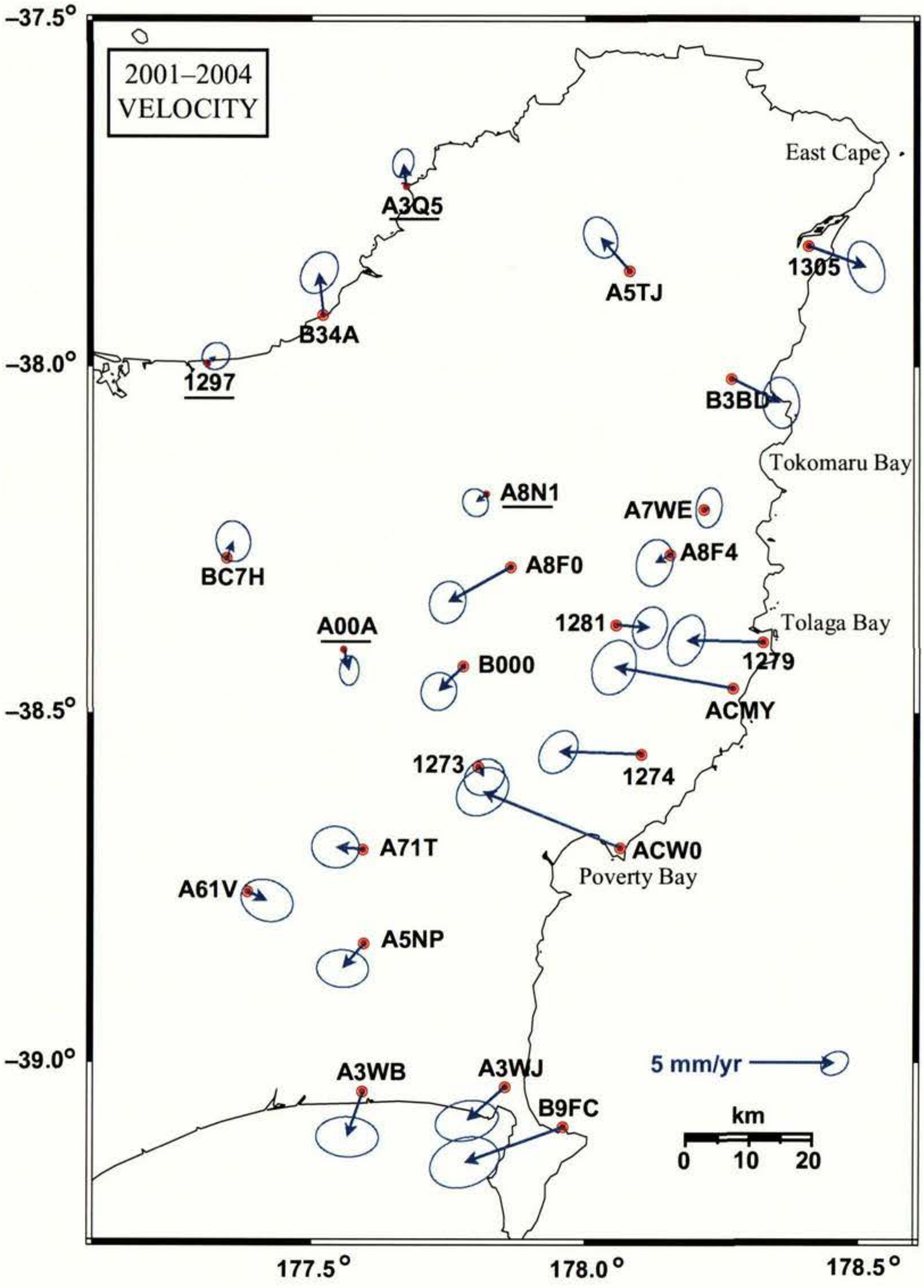

Figure 4.4: **Velocity solutions for Raukumara Peninsula GPS sites between 2001 and 2004**. The velocity solution is plotted relative to western Raukumara Peninsula stations 1297, A3Q5, A8N1, and A00A (underlined). A rotation between surveys has been calculated and removed. Rotation of the second survey coordinates (2004) $= -0.0452\ \mu$rad. Uncertainty ellipses are 95% conf.

The greater magnitude of westward displacement observed for the east coast sites between Poverty Bay and Tolaga Bay during the 2001–2004 epoch than the 1995–2001 epoch, despite the Gisborne 2002 slow slip event, is indicative of more than one slow slip event having occurred beneath the region between 1995 and 2001. If other events with eastward displacements of similar magnitude and location to the Gisborne 2002 slow slip did occur between 1995 and 2001, they may have cancelled out a large portion of the long term westward motion of those sites, resulting in the observed change in magnitude and direction of displacements. The variability in the direction of displacements for coastal campaign sites ACW0 and 1274 between epochs is in agreement with evidence from the ACW0 station position time series that westward motion of that site slowed between 1997 and 2001. These observations fit with our calculated recurrence interval for slow slip in the Gisborne region of 2–4 years.

4.2 Raukumara Peninsula strain studies

Árnadóttir et al. (1999) studied strain over the Raukumara Peninsula using triangulation surveys from 1926 and 1976 and campaign GPS data from 1995. They interpreted large variations in strain rate and orientation over the Raukumara Peninsula between the 1920s and 1995 to be the result of aseismic slip over the entire shallow subduction interface to 30 km depth, sometime between 1926 and 1976. We re-evaluate this finding based on strain calculations using all available campaign GPS data. We base our interpretations on regional deformation patterns shown by the displacement and velocity fields, and knowledge that slow slip events do occur intermittently on small sections of the deeper Hikurangi subduction interface.

4.2.1 Árnadóttir et al.'s (1999) strain study

Árnadóttir et al. (1999) computed strain with respect to a coordinate system defined by a positive x-axis pointing north and a positive y-axis pointing east. The components of displacement are u and v and $\dot{u}$ and $\dot{v}$ are the corresponding velocities. The shear strain rate components are

$$\dot{\gamma}_1 = \frac{\delta \dot{u}}{\delta x} - \frac{\delta \dot{v}}{\delta y}, \tag{4.1}$$

and

$$\dot{\gamma}_2 = \frac{\delta \dot{u}}{\delta y} - \frac{\delta \dot{v}}{\delta x}. \tag{4.2}$$

Subregion	Epoch	SEUW	Dil. rate (d) ppm/yr	$\dot{\gamma}_1$ ppm/yr	$\dot{\gamma}_2$ ppm/yr	Correlations d–γ_1	d–γ_2	γ_1–γ_2	Max ext. rate ppm/yr	β (°)	$\dot{\Gamma}$ ppm/yr
WR^i	1925–1995			−0.03±0.03	−0.05±0.03					119±15	0.06±0.04
WR	1995–2001	2.46	−0.013±0.039	0.023±0.039	0.044±0.039	−0.086	−0.456	−0.093	0.018±0.020	31.2±9.7	0.050±0.039
WR	1995–2004	2.38	−0.014±0.033	0.025±0.033	0.043±0.033	0.048	−0.266	−0.091	0.018±0.019	30.0±8.2	0.050±0.031
ER^i	1925–1995			0.09±0.04	−0.16±0.04					149±6	0.18±0.04
ER	1995–2001	5.45	0.095±0.071	−0.017±0.071	0.021±0.071	0.088	−0.226	−0.021	0.061±0.044	64.4±13.3	0.027±0.071
ER	1995–2004	5.80	0.070±0.064	−0.022±0.064	0.031±0.064	0.031	−0.201	−0.028	0.054±0.041	62.5±8.3	0.038±0.064
GIS^i	1925–1995			0.01±0.03	−0.08±0.03					139±11	0.08±0.03
GIS	1995–2001	1.71	−0.057±0.031	0.074±0.031	−0.028±0.031	0.180	−0.247	−0.024	0.011±0.022	10.2±6.5	0.079±0.031
GIS	1995–2004	3.65	−0.067±0.040	0.098±0.040	0.053±0.040	0.212	−0.308	−0.050	0.022±0.029	14.1±3.0	0.111±0.040
AP^i	1925–1976			−0.05±0.05	−0.22±0.05					129±7	0.22±0.05
AP^i	1976–1995			0.13±0.07	0.07±0.06					13±12	0.14±0.05
AP	1995–2001	4.84	0.029±0.077	−0.031±0.077	0.083±0.073	0.711	0.065	0.012	0.059±0.048	55.1±5.3	0.089±0.073
AP	1995–2004	5.71	0.015±0.074	−0.002±0.080	0.057±0.069	0.716	0.183	0.048	0.036±0.057	45.8±6.8	0.057±0.069

Table 4.1: **Strain rate results for the Raukumara Peninsula from 1925 to 2004**. Subregions of the Raukumara Peninsula (Figure 4.5) used to calculate strain were based on regions chosen by Árnadóttir et al. (1999) in their study; WR—Western Raukumara Peninsula, ER—Eastern Raukumara Peninsula, GIS—Gisborne Region, AP—Across Raukumara Peninsula. 95.0% a posteriori confidence limits are given for the 1995–2004 data. ^{i} indicates data from Árnadóttir et al. (1999) with 68.0% a posteriori confidence limits. SEUW is the standard error of unit weight of the fit of the data to the model. The model assumes uniform strain rate of the region being analysed; large SEUW values indicate that the data don't support this assumption. β is the azimuth of maximum extension. $\dot{\Gamma}$ is the maximum shear strain rate.

When positive, $\dot{\gamma}_1$ describes right-lateral shear along a northeast–southwest direction or north–south relative extension, and $\dot{\gamma}_2$ describes right-lateral shear in an east–west direction or northeast–southwest relative extension.

Árnadóttir et al. (1999) use the convention of plotting shear strain as the magnitude of the maximum shear strain rate

$$\dot{\Gamma} = \sqrt{\dot{\gamma}_1^2 + \dot{\gamma}_2^2}, \tag{4.3}$$

and the azimuth of the principal strain axis of relative extension

$$\beta = \frac{1}{2} tan^{-1} \frac{\dot{\gamma}_2}{\dot{\gamma}_1}, \tag{4.4}$$

which is perpendicular to the principal axis of contraction.

The ADJCOORD (Crook, 1992) programme used by Árnadóttir et al. (1999) to calculate strain rates over the Raukumara Peninsula does not require an exact repetition of observations in successive geodetic surveys. Thus, they were able to combine triangulation data from the 1920s and 1970s with GPS data collected in 1995. As triangulation data lack information on scale and orientation, Árnadóttir et al. (1999) were only able to estimate shear strains but not dilatation or rigid body rotation. Station coordinates and strain parameters are adjusted simultaneously until the best fit is obtained through least squares regression. Strain is assumed to vary linearly in time. Error in strain is obtained directly from the variance-covariance matrix produced during least-squares regression (Crook, 1992). Further details about ADJCOORD are found in Darby and Meertens (1995).

Árnadóttir et al. (1999) divided the Raukumara Peninsula into subnetworks to examine spatial and temporal variations in strain rate (Figures 4.5 and 4.6). A swath across the Raukumara Peninsula, where all three surveys overlapped, was studied for temporal variation in strain. Strain calculations assume that strain is uniform in space and time within each domain and within the time period between surveys.

Árnadóttir et al. (1999) found that strain varied both spatially and temporally across the Raukumara Peninsula (Table 4.1). Spatially the shear strain rate was highest in the eastern Raukumara subregion (ER) and lowest in the western Raukumara subregion (WR). Their temporal study of strain across the Raukumara Peninsula subregion (AP) showed strain rates to be very similar between epochs, but a significant difference in the azimuth of maximum extension was found between survey epochs. This variation was much larger than the uncertainties in the estimated strains.

Modeling of slip on the subduction interface, and studies of large regional earthquakes, led

Árnadóttir et al. (1999) to conclude that the temporal variation in strain across the Raukumara Peninsula could not be explained solely by coseismic and/or post-seismic effects following large Hawke Bay and Wairoa earthquakes that occurred in the early 1930s (c.f. Walcott, 1978). That is, the combined co- and post-seismic effects of the two earthquakes were too small by an order of magnitude to account for the temporal variation in strain observed between the 1925–1976 and 1976–1995 epochs for the Raukumara Peninsula. They also found that the stress change on the subduction interface resulting from the 1930s earthquakes alone was not capable of triggering two M_w 7.0 earthquakes which occurred in 1947 on the shallow subduction interface (0–15 km) offshore of the Raukumara Peninsula, and these 1947 earthquakes did not involve enough coseismic slip to fully explain the observed variation in strain.

Árnadóttir et al. (1999) speculated that elastic effects from the 1930s earthquakes initiated subsequent, progressive aseismic failure of the subduction interface, leading in turn to seismic failure of the shallow interface in the 1947 earthquakes. They proposed that strain rates observed for the 1925–1976 epoch required aseismic slip occurring above 30 km depth on the whole seismogenic portion of the subduction interface and that the plate boundary locked again between 15–30 km depth for the 1976–1995 epoch. The Árnadóttir et al. (1999) model requires a minimum dip slip rate of 40 mm yr^{-1} over an area approximately 145 km $\times$ 200 km; this gives a moment rate of 3.5×10^{19} N m yr^{-1}, equivalent to a net moment between 1925 and 1976 of 175×10^{19} N m (Table 4.2).

4.2.2 Strain estimates using recent campaign GPS data

Using recent campaign GPS data sets, we calculated the strain over similar subregions to those chosen by Árnadóttir et al. (1999), closely replicating their study (Figures 4.5 and 4.6). It is interesting to compare the Árnadóttir et al. (1999) results for the 1925–1976 period, when slow slip is hypothesised, with our results for a period when slow slip is known to have occurred. This may help us decide whether or not the strain variations that Árnadóttir et al. (1999) observed for the Raukumara Peninsula were due to an event similar to the Gisborne 2002 slow slip.

We have produced strain solutions to study both spatial and temporal variation in strain over the Raukumara Peninsula. The results are shown in Table 4.1 and Figure 4.7, together with values from Árnadóttir et al. (1999). The values in Table 4.1 were calculated in ADJCOORD assuming uniform strain across each defined subregion. The confidence limits are a posteriori

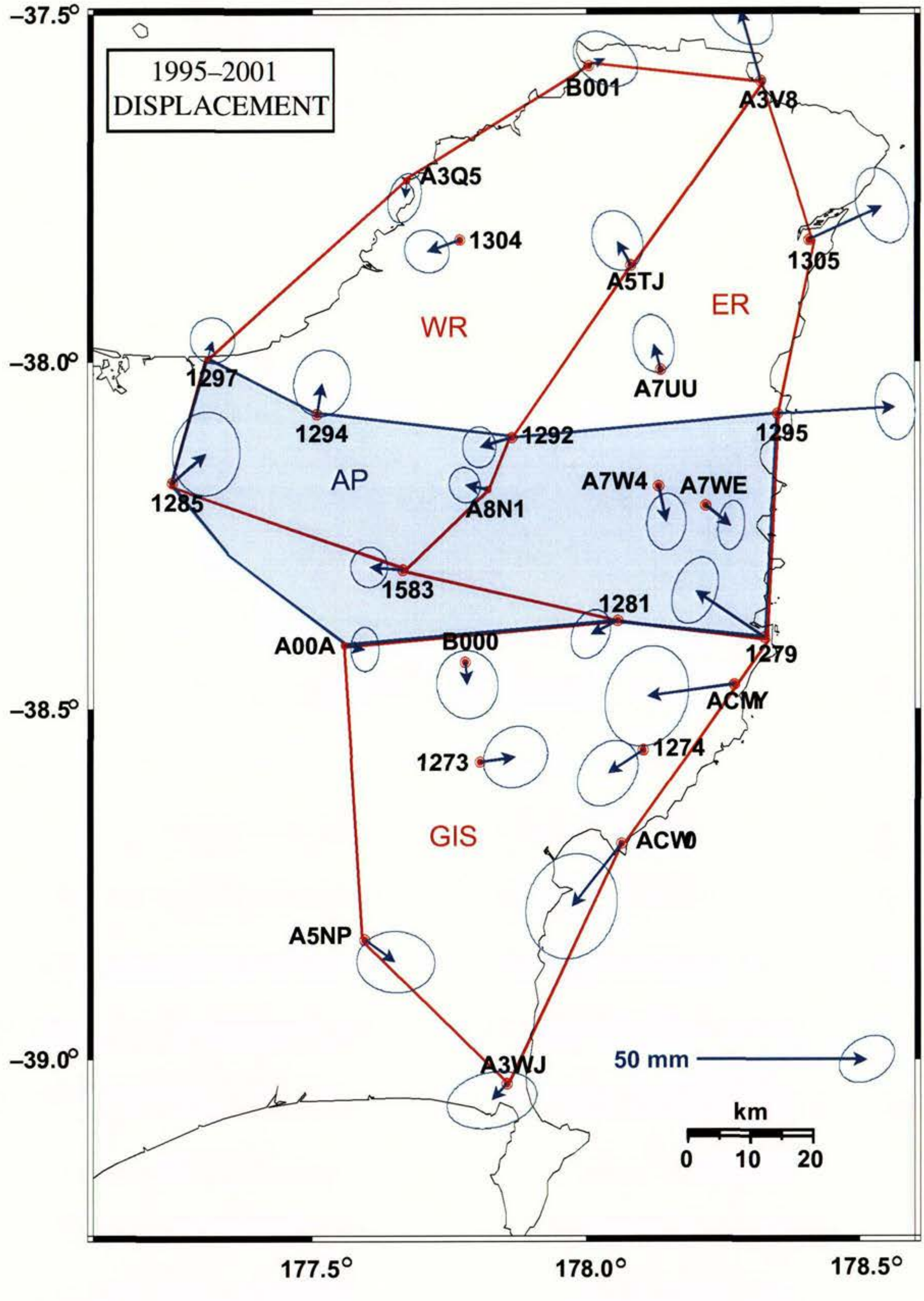

Figure 4.5: **Subregions of the Raukumara Peninsula used to calculate strain**, based on regions chosen by Árnadóttir et al. (1999). WR—Western Raukumara Peninsula, ER—Eastern Raukumara Peninsula, GIS—Gisborne Region, AP—Across Peninsula. Displacements calculated with respect to stations 1297, A3Q5, A8N1, and A00A (underlined) for the period 1995–2001 are shown. Rotation = $-0.4949\ \mu$rad. Uncertainty ellipses are 95% conf.

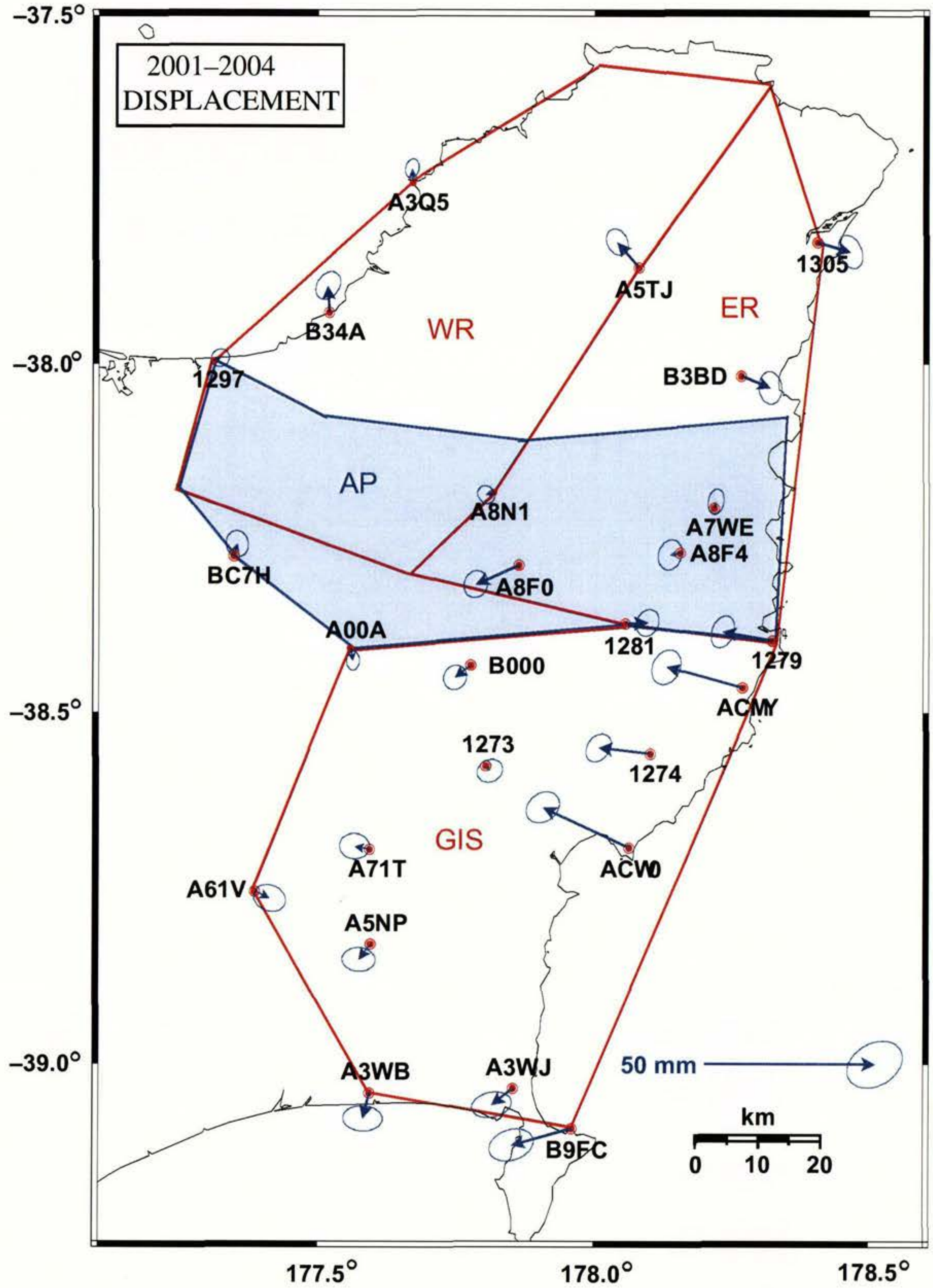

Figure 4.6: **Subregions of the Raukumara Peninsula used to calculate strain**, based on regions chosen by Árnadóttir et al. (1999); WR—Western Raukumara Peninsula, ER—Eastern Raukumara Peninsula, GIS—Gisborne Region, AP—Across Peninsula. Displacements calculated with respect to stations 1297, A3Q5, A8N1, and A00A (underlined) for the period 2001–2004 are shown. Rotation = -0.1564 μrad. Uncertainty ellipses are 95% conf.

and take into account the quality of fit of the strain model to the displacements.

Comparing the displacement fields for the Raukumara Peninsula (Figures 4.5 and 4.6) with the subregions for strain analysis chosen by Árnadóttir et al. (1999) helped us interpret the results of our spatial strain study presented in Table 4.1 using subregions eastern and western Raukumara and Gisborne. The ADJCOORD model used to calculate the strain results assumes uniform strain across each defined subregion, however the displacement fields show the deformation within each subregion to be spatially highly variable. The standard error of unit weight (SEUW) shown in (Table 4.1) gives the fit of the data to the model. Large SEUW values such as for the eastern Raukumara region (ER) indicate that the displacement data do not support the uniform strain assumption. Indeed, the displacement data shows that the large dilatation for ER (Table 4.1) is caused by the rapid eastward movement of coastal stations compared to relatively static inland stations in that region (Figure 4.5). Similarly, the compressional signal observed for the Gisborne (GIS) region is due to the westward motion of the coastal sites in that region (Figure 4.5).

To study temporal variation in strain over the Raukumara Peninsula we have generated two strain solutions for each subregion. The first solution (1995–2001) evaluates the strain over the Raukumara Peninsula prior to the Gisborne 2002 slow slip event. The second solution included the 2004 data to allow us to evaluate what effect the slow slip had on regional strain. Our results are internally consistent, with little variation between the 1995–2001 and 1995–2004 solutions. However these results do vary significantly from those of Árnadóttir et al. (1999), even considering the larger errors in their solution. In particular the azimuth of the principal strain axis of relative extension (β) is significantly different from both their 1926–1976 and 1976–1995 strain solutions. Results from the temporal (Across Peninsula—AP) study differ significantly, in magnitude and in direction of relative extension, from those of Árnadóttir et al. (1999). Large SEUW values for the AP region indicate that strain is not uniform in that region. There was little change in the magnitude or direction of strain across the Raukumara Peninsula between the 1995–2001 and 2001–2004 data sets, when the Gisborne 2002 slow slip event is known to have taken place.

Figure 4.7 shows that there is no significant change in shear strain rate either across the Raukumara Peninsula or in the Gisborne region between 2001 and 2004, over a period when slow slip events are known to displace continuous GPS sites in the Gisborne region. For the 'Across Peninsula' region, there is a small but not statistically significant clockwise rotation of the 1995–2001 and 1995–2004 strain-rate azimuths as compared to 1976–1995. However, these three directions all differ significantly from the 1925–1976 direction (Figure 4.7).

Our study and that of Árnadóttir et al. (1999) both show that strain does vary significantly in space and time over the Raukumara Peninsula. However, our results show that the Gisborne 2002 slow slip event did not have a significant effect on the regional strain picture for the Raukumara Peninsula. In comparison to our slow slip model (Figure 3.4), in which a small (25 km$\times$60 km) portion of the subduction interface offshore of Gisborne slips, the Árnadóttir et al. (1999) model requires slip on the whole seismogenic portion of the subduction interface above 30 km, along the length of the Raukumara Peninsula. While much larger than our modeled slip plane for the 2002 Gisborne slow slip event, the area of slip proposed by Árnadóttir et al. (1999) is not unusual in the context of the Pacific rim (Table 4.2). However, the net moment release required by their model is anomalously large.

In order to compare slow slip events modeled for the northern Hikurangi subduction margin with events recorded elsewhere, we calculate the stress drop of selected events based on inferred modeled slip parameters. Stress drop (Table 4.2) is calculated using the equation given by Lay and Wallace (1995) for dip slip faulting, $\Delta\sigma \approx \mu\bar{D}/w$, assuming a Poisson ratio of 0.25. Stress drop is independent of along-strike length for a dip slip fault, enabling us to compare the Gisborne 2002 slow slip with events elsewhere without needing a precise estimate of the fault length. Table 4.2 shows that while the preferred model slip plane of the Gisborne 2002 slow slip event has a low moment release, it has a high stress drop compared to other Pacific rim events. The stress drop of the Árnadóttir et al. (1999) slip model is substantially higher, although in that case the width of the slip plane is poorly constrained.

The 50 year hiatus between triangulation surveys on the Raukumara Peninsula and the occurrence of five major earthquakes in that time makes it impossible to estimate a specific time window for a regionally extensive slow slip event on the northern Hikurangi subduction interface, such as that modeled by Árnadóttir et al. (1999). Recent continuous GPS records show, however, that smaller slow slip events currently occur episodically on the subduction interface beneath the Gisborne region. We find it unlikely that the anomalously large moment release required by the Árnadóttir et al. (1999) model would be the result of one slow slip event beneath the Raukumara Peninsula. Rather, it is more likely that a number of slow slip events occurred between the 1920s and 1970s, cumulatively contributing to the observed change in strain across the Raukumara Peninsula.

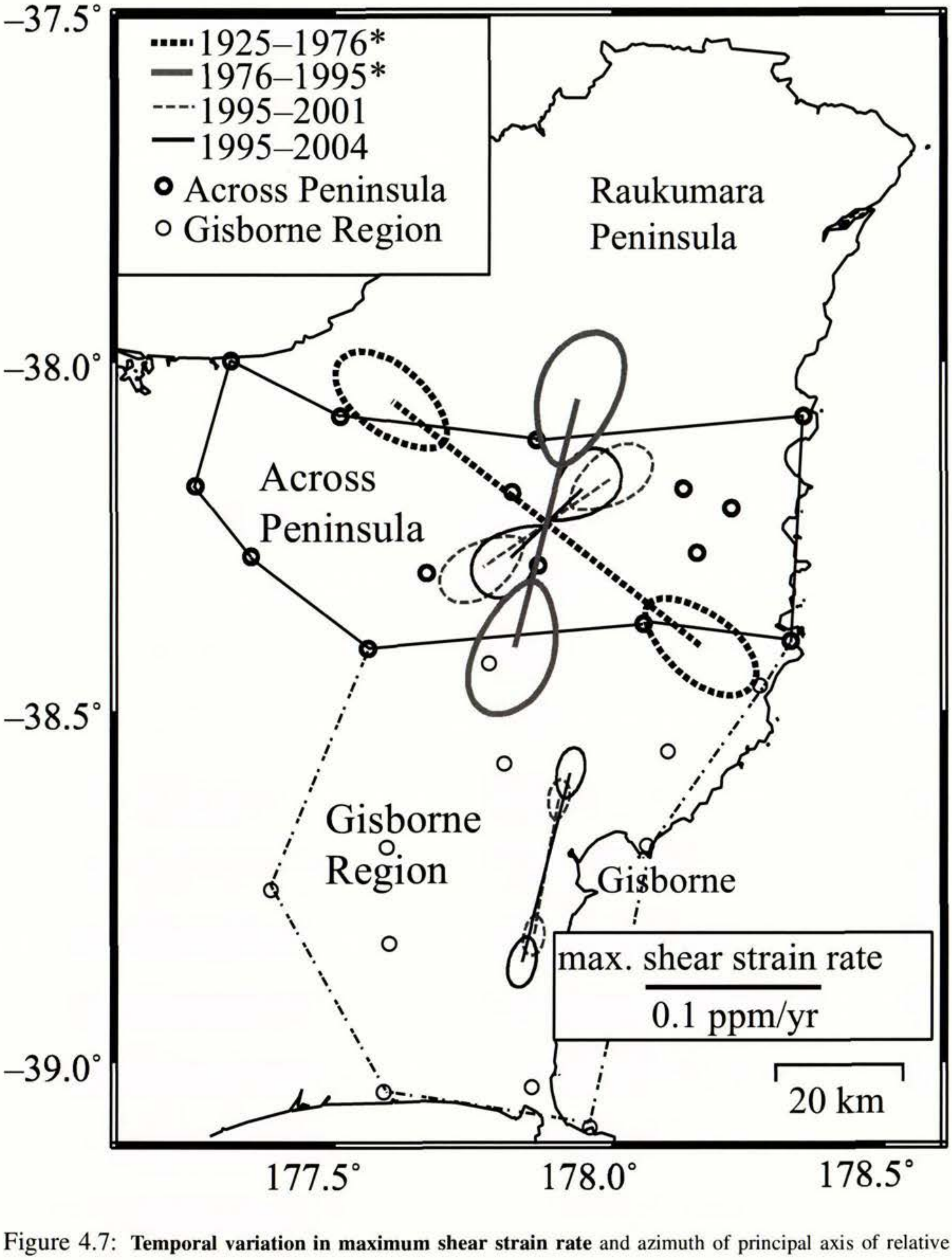

Figure 4.7: **Temporal variation in maximum shear strain rate** and azimuth of principal axis of relative extension across the Raukumara Peninsula(from triangulation and GPS data between 1925 and 2004); * data from Árnadóttir et al. (1999), and for the Gisborne region (from 1995–2004 GPS data). Uncertainties are 2-D 68% confidence regions taking account of random error and the misfit to a uniform strain model.

Slow slip event	Year	Slip ($\bar{D}$) (m)	Length (L) (km)	Width (w) (km)	Moment release (N m)	Stress drop (Pa)
Bungo Channel[i]	1997	0.6	230	100	8×10^{19}	1.8×10^{5}
Cascadia[ii]	1999	0.02	300	50	1.4×10^{19}	1.2×10^{4}
Guerrero Gap[iii]	2001	0.1	550	250	40×10^{19}	1.4×10^{4}
Raukumara Pen.[iv]	1920s–1970s	2	200	145[iv]	175×10^{19}	4.1×10^{5}
Gisborne[v]	2002	0.18	60	25	0.81×10^{19}	2.4×10^{5}

Table 4.2: **Area and moment release of selected Pacific rim slow slip events**

[i] Miyazaki et al. (2003); slip is a maximum value

[ii] Dragert et al. (2001); slip is an average value

[iii] Kostoglodov et al. (2003); slip is an average value

[iv] Árnadóttir et al. (1999); slip is a minimum value and the updip width of the slip plane is not constrained

[v] This study; Model B slip parameters in which the north–south extent of the slip plane is not constrained by the Raukumara Peninsula geodetic data.

4.3 Summary

Total displacement and velocity fields for the Raukumara Peninsula show three distinct regions of deformation. Coastal sites between Poverty Bay and Tolaga Bay show rapid westward displacement. The greater magnitude of westward displacement observed for the east coast sites between Poverty Bay and Tolaga Bay during the 2001–2004 epoch than the 1995–2001 epoch, despite the Gisborne 2002 slow slip event, is indicative of more than one slow slip event having occurred beneath the region between 1995 and 2001. Variability in the direction of displacements for coastal campaign sites ACW0 and 1274 between epochs supports this observation. The interpretation that more than one slow slip event occurred in the Gisborne region between 1995 and 2001 fits with our calculated recurrence interval for slow slip of 2–4 years. The generally rapid westward displacement of coastal sites between Poverty Bay and Tolaga Bay reflects a zone of elevated inter-seismic coupling on the subduction interface below this region and is consistent with to the the westward subduction of the Pacific plate under the North Island (Wallace et al., 2004). Central and western Raukumara campaign sites have much smaller displacements than the east coast sites and show no clear pattern of deformation.

Our study of strain across the Raukumara Peninsula and that of Árnadóttir et al. (1999) both show that strain varies significantly in space and time over the region, but there is little variation in strain due to the Gisborne 2002 slow slip event. Árnadóttir et al. (1999) interpret the variation in strain, which occurred across the Raukumara Peninsula between the 1920s and

1970s, to be the result of slow slip extending over the whole of the seismogenic portion of the subduction interface. This slip region is significantly larger than our preferred model slip plane for the 2002 Gisborne slow slip event, and while the area of slip is not unusual for Pacific rim subduction zones, the corresponding moment release required is much larger than elsewhere. Slow slip beneath the Raukumara Peninsula currently occurs in episodic events causing slip on restricted regions at the base of the seismogenic zone on the subduction interface offshore of Gisborne. That slow slip events on the northern Hikurangi margin are currently localised, is supported by our finding that the Gisborne 2002 slow slip event had no significant effect on regional deformation patterns.

Chapter 5

Tremor

The linking of slow slip events to unique seismic tremor signals on the Japan and Cascadia subduction margins in the past three years has been a significant advance in the study of slow slip and seismogenesis on subduction margins. We analyse continuous broadband seismic data from the Hikurangi subduction margin to investigate the occurrence of tremor in association with the two largest slow slip events from the Gisborne region, the 2002 and 2004 events. These preliminary studies suggest that both slow slip events were accompanied by seismic tremor; however, a more detailed study is necessary to determine whether there is a temporal and spatial correlation between the slow slip and tremor. Here we present the methods and results of our preliminary studies; the study of tremor on the Hikurangi subduction margin is an important area for future work.

5.1 Analysis method

Following methods used by scientists in Cascadia (McCausland, W., Szeliga, W., Rogers, G., 2004, pers. comm.), we have analysed seismic data from continuously recording broadband seismometers on the Raukumara Peninsula, Hawke's Bay and Waikato to determine whether tremor occurred during the 2002 and 2004 Gisborne slow slip events on the Hikurangi subduction interface.

We obtained miniseed format continuous broadband seismic data directly from GeoNet in 24 hour-long files. This data was converted into SAC (Seismic Analysis Code)[1] format, which were then cut into hour-long blocks for analysis. For simplicity, we analysed the vertical component of the broadband data only. However, it should be noted that Rogers and Dragert

[1] http://www.llnl.gov/sac/

(2003) found the Cascadia tremor signals to be visible on all components but strongest on the horizontal.

We use data from broadband seismic station TOZ in the Waikato as a reference site for the stations on the eastern side of the Central Volcanic Region (CVR). Small seismic signals would not propagate across the CVR (Smith, E., 2004, pers. comm.), therefore signals such as low frequency seismic tremor generated locally on the East Coast would not be recorded on station TOZ, whereas large earthquakes would be. This is a good way of distinguishing between local and far-field seismic sources.

Tremor recorded on the Cascadia margin has its energy mainly in the 1–5 Hz range. When recorded on a single station, the tremor signal is unremarkable as it falls within the frequency range of environmental (due to weather or ocean tides) and cultural noise. In Cascadia, as in Japan (Rogers and Dragert, 2003; Obara, 2002) the tremor was observed simultaneously at many stations as far as 300 km apart across the seismic network allowing such noise to be discounted.

In October 2002 only one broadband seismic station (KNZ) was running continuously in the immediate Gisborne region (Figure 5.1). We made use of data from three more distant stations, URZ, PWZ and TOZ; however, due to the low signal-to-noise ratio on station KNZ in particular, and the small number of stations, it was difficult to identify possible tremor episodes across the seismic network for the 2002 slow slip event. As detailed analysis of hour-long blocks of the 2002 broadband seismic data proved unsuccessful, we generated RSAM (Real-time Seismic-Amplitude Measurement) plots from three months spanning the 12-day slow slip event for stations KNZ and PWZ to gain an overview of seismicity during that time period (Figure 5.2). RSAM, which is typically used in the monitoring of active volcanoes (e.g. Sherburn et al., 1996), computes the average amplitude of ground shaking caused by earthquakes and tremor over a specified time interval.

The RSAM plot for stations KNZ and PWZ shows an increase in seismic energy relative to background levels on these stations around the time of the Gisborne 2002 slow slip event (Figure 5.2). Such an increase in seismic energy could be attributed to cultural or environmental noise, so to eliminate these factors we would need to see the tremor signal on multiple stations in the same time window. It would also be important to study a much longer time period of continuous seismic data to see whether such increases in seismic energy also occur at times when no slow slip is recorded.

At the time of the November 2004 slow slip event five broadband seismometers were con-

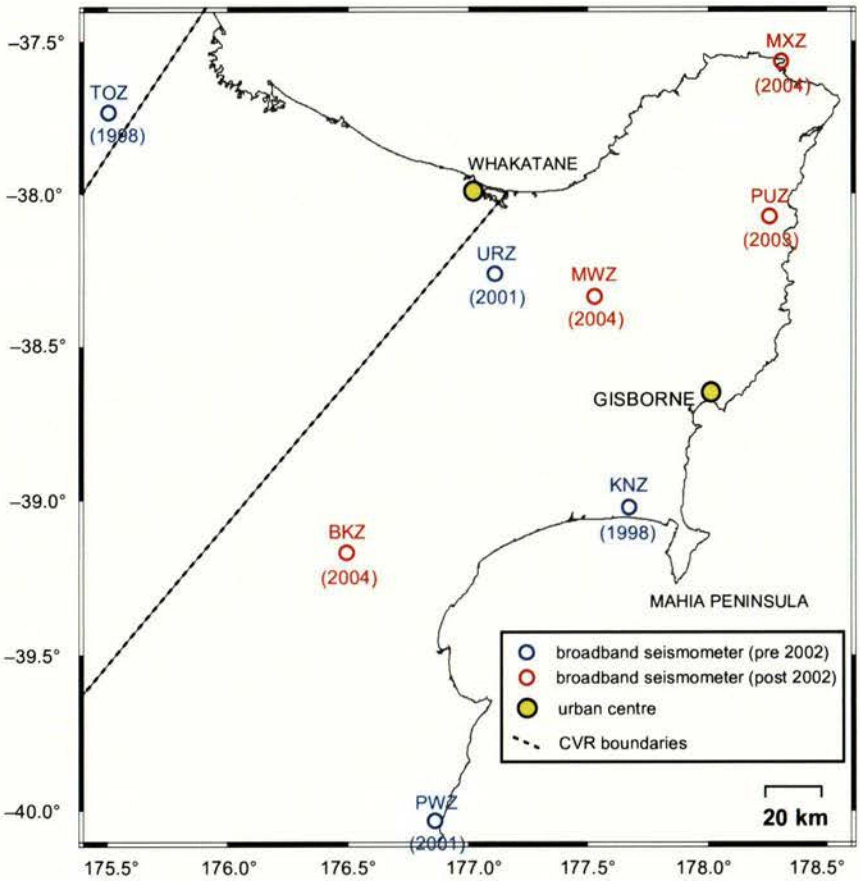

Figure 5.1: **Continuous broadband seismic stations used for tremor studies.** The year of installation is shown beneath the station name; stations installed prior to 2002 are blue; stations installed after 2002 are red. Approximate eastern and western boundaries of the CVR are dashed lines.

tinuously recording on the Raukumara Peninsula itself (Figure 5.1). This greatly enhanced the possibility of identifying tremor. For the 2004 slow slip event, we analysed twenty-five days of broadband seismic data from seven regional stations, starting five days prior to the event.

Following the method of McCausland (pers. comm., 2004), University of Washington, we read the hour-long files for all available stations into SAC. We removed the mean and then used the SAC glitch removal command, 'rglitches', to remove station glitches and generate a continuous data file. We strongly bandpass filtered the data between 1 and 6 Hz (Figure 5.3A), then took the absolute value of the bandpassed file and incrementally smoothed and decimated the signal down to 0.625 Hz. The resulting file is an 'envelope' of the bandpassed file which can be rapidly scanned by eye for signals which occur on many stations in the same time window (Figure 5.3B). Signals of interest can be more closely examined in SAC by reducing the time window — Figure 5.4 shows a ten minute time window. This method was more successful for the 2004 slow slip event as there were four broadband seismic stations

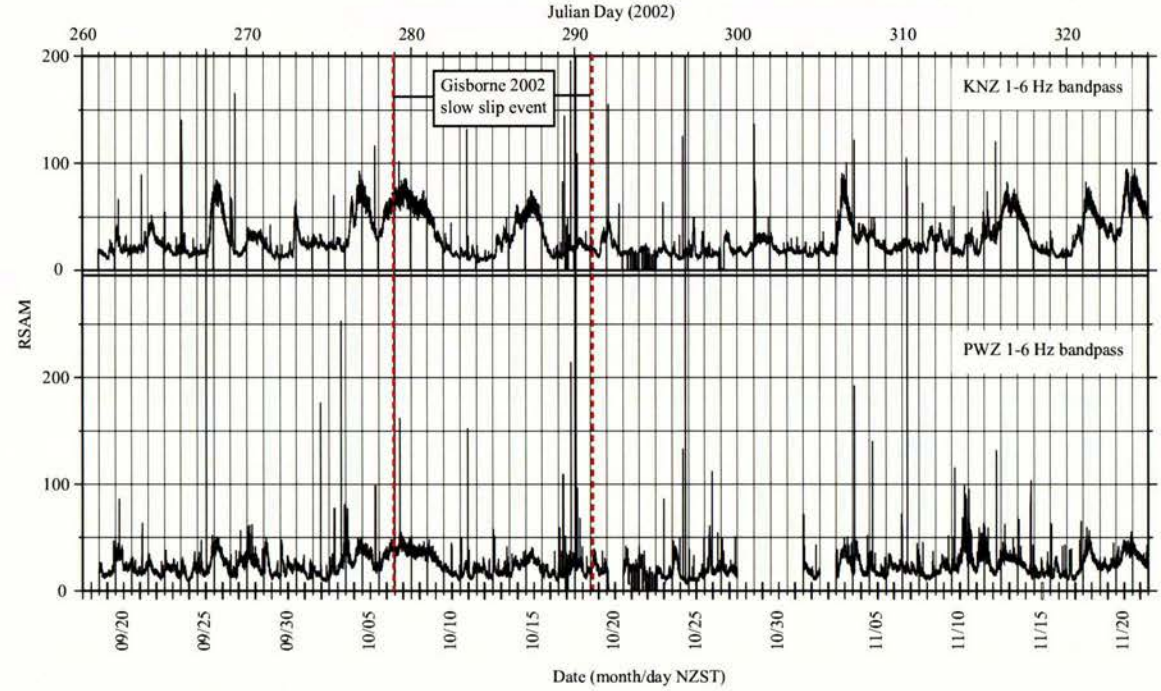

Figure 5.2: **RSAM of low frequency (1–6 Hz) bandpass broadband sesimic data** from stations KNZ and PWZ for three months (Sept–Nov) spanning the 12-day Gisborne 2002 slow slip event. Red dashed lines indicate the time window of the slow slip event.

continuously recording within a 100 km radius of Gisborne.

If a possible tremor signal is located during the visual analysis of envelope files, we refer back to the original, and bandpass filtered SAC files to double-check the signal. Figures 5.3 and 5.4 show a tremor signal from the time window of the 2004 slow slip events. A second example is shown in Figure 5.5 with a ten minute window only. The characteristic tremor signal, which we look for when scanning the envelope files, has a gradual onset, and forms a broadly symmetrical envelope. The tremor has little energy above 6 Hz as illustrated by Figure 5.4, C&D. Large regional earthquakes (Figure 5.6) can be clearly identified due to the instantaneous increase in energy in the envelope file and the asymmetrical shape of the envelope. Earthquakes also have a significant component of energy at frequencies higher than 6 Hz (Figure 5.6, C&D).

These initial results from the 2004 slow slip event are encouraging. Studies of longer periods of seismic data will help to identify whether the tremor signals observed during this event are temporally associated with the slow slip, or whether they are pervasive in the broadband seismic record. Ideally, a study of tremor for the Raukumara region should look at at least three years worth of broadband seismic data to clearly identify patterns of tremor and its cor-

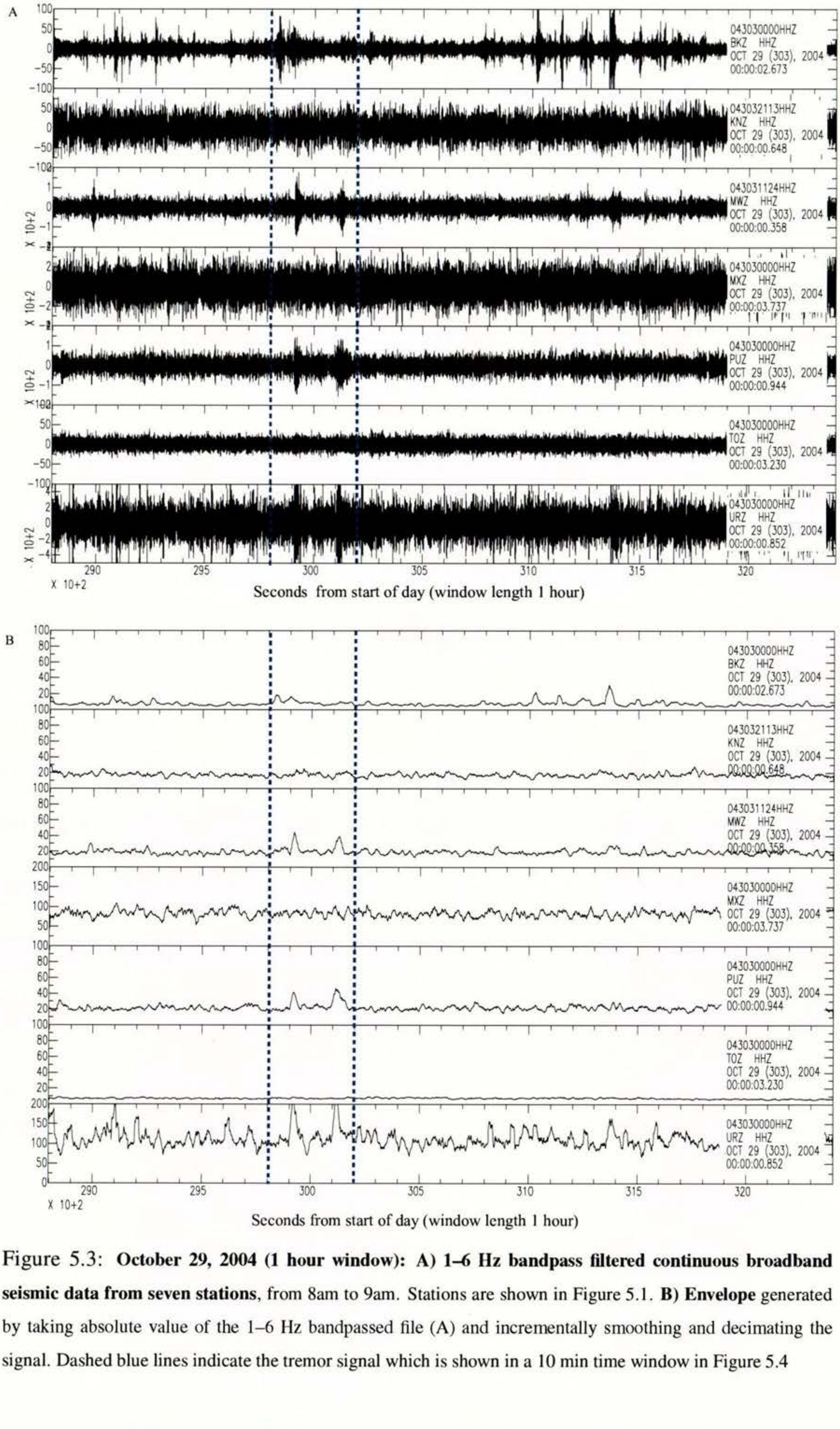

Figure 5.3: **October 29, 2004 (1 hour window): A) 1–6 Hz bandpass filtered continuous broadband seismic data from seven stations**, from 8am to 9am. Stations are shown in Figure 5.1. **B) Envelope** generated by taking absolute value of the 1–6 Hz bandpassed file (A) and incrementally smoothing and decimating the signal. Dashed blue lines indicate the tremor signal which is shown in a 10 min time window in Figure 5.4

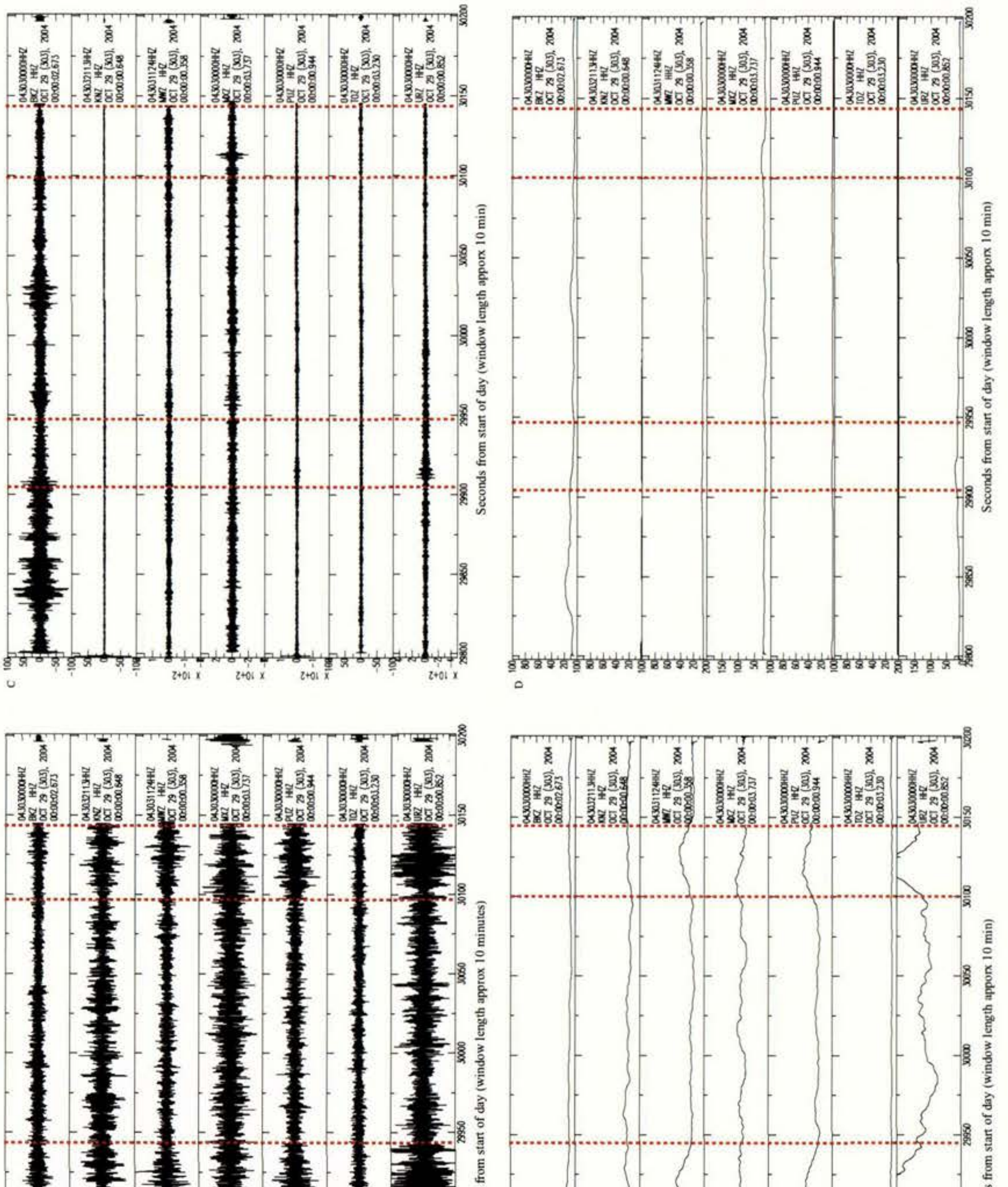

Figure 5.4: **October 29, 2004 (10 minute window): A) 1–6 Hz bandpass filtered continuous broadband seismic data** for 10 minutes between 8 am and 9 am, expanded from Figure 5.3. Stations are shown in Figure 5.1. **B) Envelope** generated by taking absolute value of the 1–6 Hz bandpassed file (A) and incrementally smoothing and decimating the signal. The tremor signal, indicated by dashed red lines, has a gradual onset and forms a broadly symmetrical envelope; it is not recorded on station TOZ. **C) 6–10 Hz bandpass; D) Envelope of 6–10 Hz bandpassed file (C)** showing little energy above 6 Hz, a characteristic of tremor. Dashed red lines indicate tremor region identified in (A) and (B).

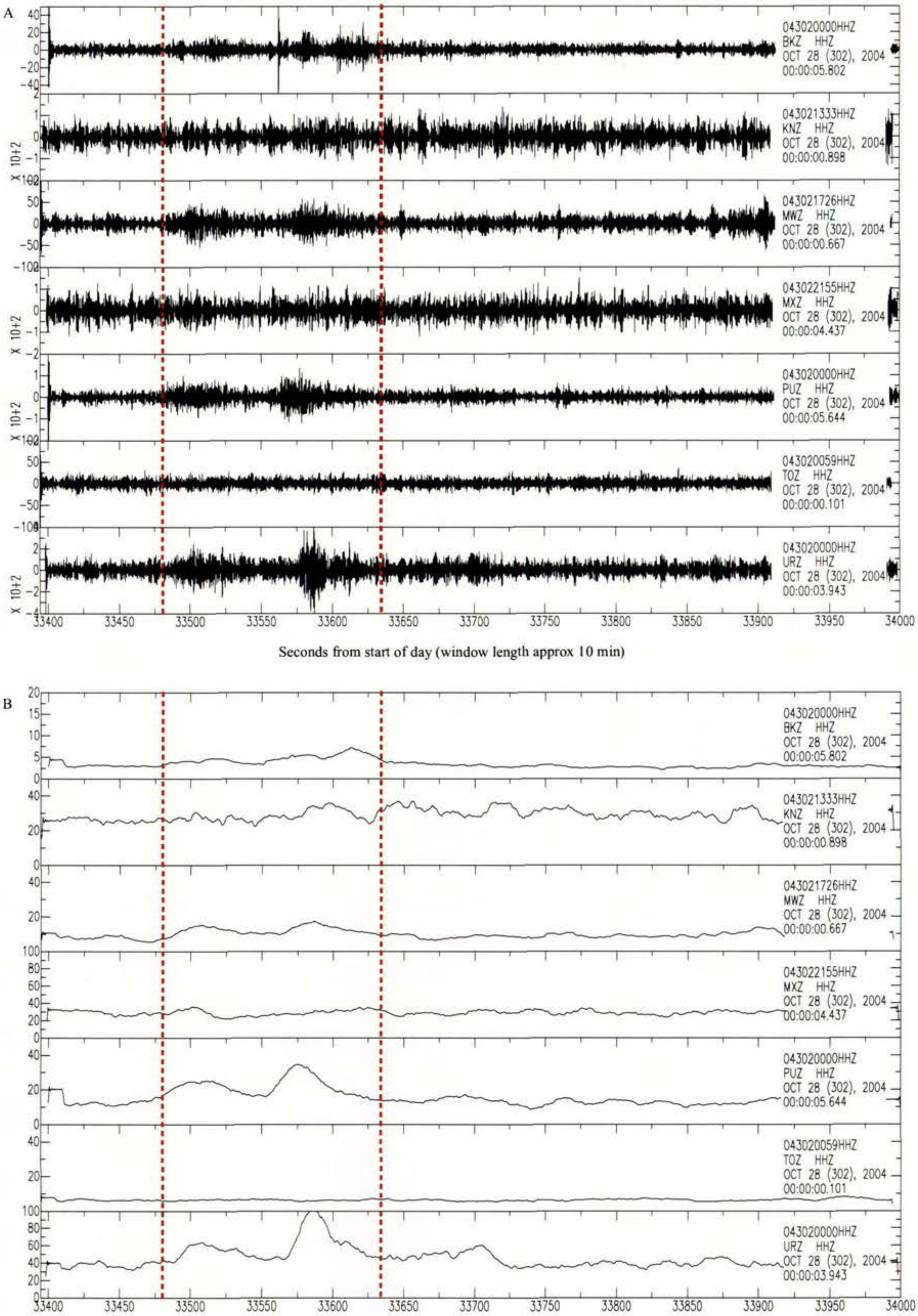

Figure 5.5: **October 28, 2004 (10 minute window): A) 1–6 Hz bandpass filtered continuous broadband seismic data** for 10 minutes. Stations are shown in Figure 5.1. **B) Envelope** generated by taking absolute value of the 1–6 Hz bandpassed file (A) and incrementally smoothing and decimating the signal. The tremor signal, indicated by dashed red lines, has a gradual onset and forms a broadly symmetrical envelope.

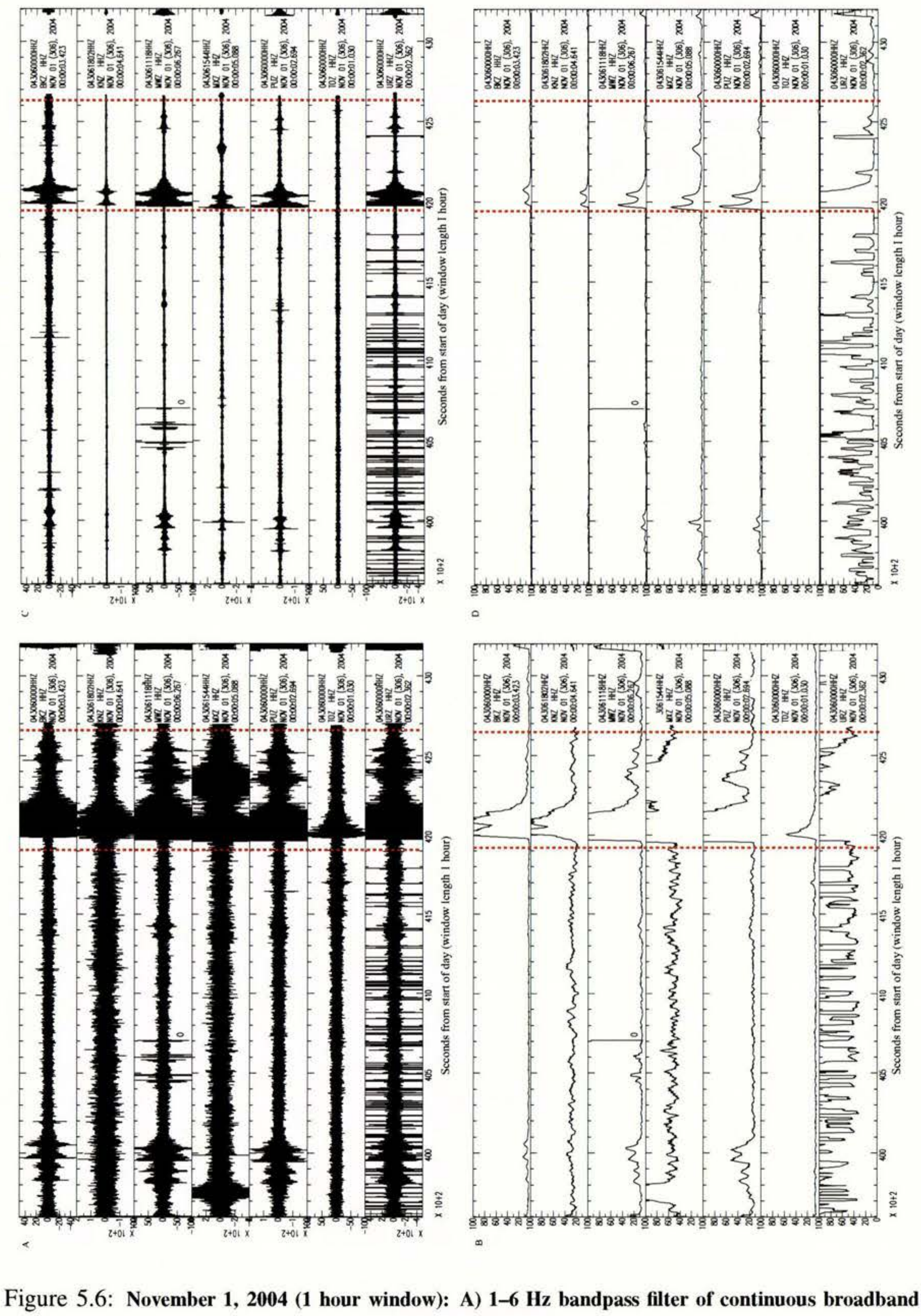

Figure 5.6: **November 1, 2004 (1 hour window): A) 1–6 Hz bandpass filter of continuous broadband seismic data from seven stations** from 11am to 12 noon. Stations are shown in Figure 5.1. **B) Envelope** generated by taking absolute value of the 1–6 Hz bandpassed file (A) and incrementally smoothing and decimating the signal. Dashed red lines indicate an earthquake, distinguished from tremor due to the instantaneous increase in energy and asymmetrical shape of the envelope. The earthquake signal is recorded on station TOZ. **C) 6–10 Hz bandpass filter. D) Envelope of 6–10 Hz bandpassed file (C)** showing that the earthquake has a significant component of energy at frequencies higher than 6 Hz.

relation to slow slip, as has been done for the Cascadia margin (e.g. Rogers and Dragert, 2003; Szeliga et al., 2004). If a clear temporal relationship with the slow slip events was established for the Hikurangi margin, it would then be important to determine the extent to which tremor occurs in the region and where it is located.

5.1.1 Attenuation

In the study of tremor, regional seismic attenuation plays an important role. In southern Cascadia, where tremor was observed over a broad area, the low attenuation of the granite basement allowed tremor signal to travel well over long distances (Szeliga, W., 2004, pers. comm.). In the Raukumara region, the eastern half of the Raukumara Peninsula north and south of Gisborne is formed by Neogene marine sedimentary rocks. These rocks are underlain by the Late Cretaceous and Early Tertiary rocks of the East Coast Allochthon (Mazengarb and Speden, 2000; Thornley, 1996). Both of these units have high seismic attenuation (Eberhart-Phillips and Chadwick, 2002). Zones of high attenuation beneath the Raukumara Peninsula are also associated with postulated fluid flow caused by dehydration of subducted slab sediment (Eberhart-Phillips and Chadwick, 2002). In the case of tremor studies on the Raukumara Peninsula, choosing a lower and more restricted range for the bandpass filter (e.g. 1–2 Hz) which would attenuate less, may maximise the potential of correlating seismic tremor with slow slip on the low density broadband network of the Raukumara Peninsula (Smith, E., 2004, pers. comm.).

5.2 Relationship between tremor and slow slip

Studies of the Cascadia episodic tremor and slip events show that the epicenters of most tremor signals are found to be confined to a limited horizontal band bounded approximately by the horizontal locations of the 30 and 45 km depth contours of the subducting plate interface (Kao et al., 2004). However, the depth range for the tremor was wide, extending from the upper crust to the subducted Juan de Fuca slab, with a peak at 25–35 km depth. Much of the episodic tremor appeared to coincide with strong seismic reflectors in the overriding plate where no local earthquakes had been observed (Kao et al., 2004).

Seismic tremor signals are interpreted to be caused by fluid flow (Obara, 2002). The observation by Kao et al. (2004) that tremor associated with slow slip is not confined to the plate interface, but is distributed throughout the volume of the upper plate, suggests that fluid reservoirs are excited by slip-generated strain resulting in tremor, rather than fluid flux exciting slip (Lowry et al., 2005). Whatever the physical relationship between tremor and slow slip, the close spatial and temporal association observed in Cascadia and Japan suggests that routine monitoring of tremor might provide a valuable means of tracking the occurrence and/or evolution of slow slip (e.g. Obara et al., 2004; Szeliga et al., 2004).

5.3 Future study

The study of tremor in association with slow slip events on the Hikurangi subduction margin is an important area for future research. While we have made some preliminary observations in this chapter our findings remain inconclusive regarding the relationships between tremor and slow slip beneath the Raukumara Peninsula. Studies of longer periods of seismic data will help to identify whether the signals observed during the 2004 slow slip event in particular are temporally associated with the event, or whether they occur more frequently. Whereas the main focus of this thesis has been geodetic evidence for slow slip on the northern Hikurangi subduction margin, the study of seismic tremor is an equally important component in unraveling the complex picture of how deformation occurs along the subduction margin.

Chapter 6

Conclusion

6.1 Summary

The aim of this project was to identify and characterise the deformation associated with slow slip events beneath the Raukumara Peninsula on the northern Hikurangi subduction margin. We have achieved this through:

1. Re-analysis of existing GPS data using modern software and uniform satellite orbits;

2. Collection and analysis of new campaign GPS data;

3. Interpretation of the combined 1995–1997–2001–2004 geodetic data set;

4. Investigation of the recent continuous broadband seismic data set.

We have processed and analysed nearly ten years of regional campaign GPS records (1995–2004) in conjunction with recent continuous GPS and broadband seismic data. We have used these data to study a slow slip event, which generated 20–30 mm of surface displacement in the Gisborne region over ten days in October 2002 (Beavan et al., 2003). This event was recorded on two local continuous GPS stations, and we have used the regional campaign data to study the spatial extent of the slow slip event and its effect on regional deformation.

Our work on station position time series from the campaign GPS data set shows that low temporal spacing of observations has aliased the Gisborne 2002 slow slip event. We find that we cannot use the campaign GPS data to constrain the spatial extent and moment release of this event. However, we can successfully use the time series to make estimates of the recurrence interval for events of similar surface displacement to that in 2002 occurring in the Gisborne region. Our calculations suggest that such events recur at 2–4 year intervals, which is in agreement with the interval between the October 2002 and recent November 2004 slow slip events in the Gisborne region. Our tentative interpretation that a slow slip event occurred in the region between 1997 and 2001 also fits this time frame.

Our preferred forward model of the Gisborne 2002 slow slip event shows it to be the result of 18 cm of slip on a plane, approximately 60 km $\times$ 25 km in area, on the subduction interface offshore of Gisborne, though the north–south extent is poorly constrained. The model slip plane coincides with the downdip limits of the seismogenic zone on the subduction interface (Reyners, 1998; Wallace et al., 2004). This finding is in agreement with model slip locations for slow slip events occurring elsewhere on the Pacific rim (e.g. Dragert et al., 2001; Obara et al., 2004; Larson et al., 2004). In order to examine any variations in inter-seismic coupling and slip deficit on the northern Hikurangi margin due to the 2002 Gisborne slow slip event, we have also added the 2004 campaign GPS data to models described by Wallace et al. (2004). This modelling shows that there was no significant change in inter-seismic coupling and slip deficit that could be directly attributed to the 2002 event. A zone of elevated coupling beneath the Tolaga Bay area of the Raukumara Peninsula persists in spite of the relative plate motion between the Australian and Pacific plates being intermittently accommodated through slow slip events on the subduction interface.

We have generated total displacement and velocity fields for the Raukumara Peninsula using the campaign GPS data set, and find that the regional deformation pattern over the Raukumara Peninsula was not noticably affected by the Gisborne 2002 slow slip event. The dominant features of regional deformation persist: the rapid westward displacement of coastal sites between Poverty Bay and Tolaga Bay reflects a zone of elevated inter-seismic coupling on the subduction interface below this region (Wallace et al., 2004); the predominantly eastward motion of northeastern Raukumara Peninsula sites is consistent with rotation of the Raukumara Peninsula away from the TVZ.

Our study of strain across the Raukumara Peninsula, combined with that of Árnadóttir et al. (1999), shows that strain varies significantly in space and time over the region, but we find that there is little discernible variation in strain over the time period of the Gisborne 2002 slow slip event. In general, our study of total displacement and velocity fields, and strain across the Raukumara Peninsula shows that the Gisborne 2002 slow slip event had little effect on regional deformation patterns.

Árnadóttir et al. (1999) interpreted large strain variations across the Raukumara Peninsula between the 1920s and 1990s to be the result of slow slip extending over the whole of the seismogenic portion of the subduction interface. This is significantly larger than our preferred model slip plane for the 2002 Gisborne slow slip event, but the area of slip is not unusual in the context of Pacific rim subduction zones (Table 4.2). In contrast, our model of slip suggests that slow slip beneath the Raukumara Peninsula currently occurs in episodic events involving local slip at the base of the seismogenic zone on the subduction interface offshore of Gisborne. This is supported by our finding that the Gisborne 2002 slow slip event had no significant effect on regional deformation patterns. It is likely that slow slip on the northern Hikurangi subduction margin occurs over a range of rupture areas, from smaller slip planes like the 2002 and 2004 events, to regionally extensive slip areas as postulated by Árnadóttir et al. (1999).

Based on methods developed by Japanese and North American scientists, we have conducted a preliminary study of the possible association of seismic tremor with the slow slip events recorded geodetically on the Raukumara Peninsula. We find evidence for tremor recorded on several stations across the Raukumara Peninsula broadband seismic network during both the 2002 and 2004 Gisborne slow slip events, but we have not rigorously examined the spatial and temporal relationships of this tremor to the slow slip events.

6.2 Future work

In this thesis, our modelling has focused on the Gisborne 2002 slow slip event. In November 2004, another slow slip event of similar ground displacement to the 2002 event was recorded in the Gisborne region. At the time of this event, three GeoNet stations were newly installed in the Raukumara network and due to their very short station position time series we were unable to confirm whether these new stations had been displaced by the 2004 event. Based on the slow slip recurrence interval of 2–4 yrs computed in this thesis, we would expect another slow slip event to occur in the Gisborne region in late 2006 or 2007. The denser, and by then well-established, network of continuous GPS receivers on the Raukumara Peninsula will allow the spatial extent and moment release of such future events to be better constrained, and the strategic location of continuous GPS receivers along the strike of the northern Hikurangi subduction margin will help establish whether slow slip events, such as that in 2002, propagate along strike on the subduction interface. Examples from Cascadia, Japan and the Guerrero Gap illustrate that slow slip events accommodate significant moment release at subduction zone plate boundaries. Quantifying the moment release contribution of slow slip on the northern Hikurangi subduction margin will help us to more accurately assess the seismic hazard posed by this subduction zone to New Zealand.

The study of tremor in association with slow slip events on the Hikurangi subduction margin is an important area for future research. While we have made some preliminary observations, our findings regarding either the temporal or spatial association of seismic tremor with slow slip remain inconclusive. Studies of longer periods of seismic data will help to identify whether the signals observed during the 2004 slow slip event in particular are temporally associated with the event. Although the main focus of this thesis has been geodetic evidence for slow slip on the northern Hikurangi subduction margin, the study of seismic tremor has been shown elsewhere to play an equally important part in unravelling the complex picture of how deformation occurs along the subduction margin.

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Appendix A

Raukumara Peninsula GPS data collection

A.1 GPS data collection

The GPS data used for this thesis were collected during four survey campaigns, in 1995, 1997, 2001 and 2004, by teams from the Department of Survey and Lands (DOSLI) and Victoria University of Wellington (VUW) and Institute of Geological and Nuclear Sciences (GNS). Here, field methodology from the 2004 GPS campaign is explained.

A.1.1 Logistics and site selection

Gisborne is the major city of the Raukumara Peninsula and was the base of operations for all the GPS campaigns. GPS receivers were commonly deployed using four-wheel drive vehicles or quad bikes. In 1995 and 2001, teams also made helicopter deployments. In 2004, our deployments were aided by fine weather, but in past years some four wheel drive access routes were treacherous or impassable in wet conditions.

In 2004 we selected sites from the Gisborne area which had been occupied three or more times in the past. We also occupied sites on the western Raukumara Peninsula to replicate the spatial extent of previous surveys. We gained the permission of the landowner for all the sites that we visited. On arrival at each site we checked sky visibility and noted any obstructions which could block the sky view of the antenna. This was especially relevant to stations near forestry blocks where the vegetation had changed since the last survey. Clear line of sight to the antenna is important as it minimises the chance of multipath errors, which occur when the

GPS signal reaches the antenna via two or more paths, or cycle slips, which can occur if the satellite signal is obstructed.

Most Raukumara Peninsula sites are marked by survey benchmarks or pins cemented into holes drilled in stable bedrock. Four meter triangulation beacons over the survey marks must be temporarily detached and moved away from the site marker to prevent signal obstruction during the survey period. These 'quadripods' are common on the Raukumara Peninsula and became an important logistical consideration as three people were usually required to resurrect them. In some cases smaller permanent beacons (Nelson beacon) are cemented into the ground at the site. In this case an antenna can be fixed directly to the beacon and no tripod is required.

A.1.2 Standard station setup using a tripod

- **Setup tripod:** An adjustable tripod is set up above the benchmark or pin, approximately levelled and fixed securely into the ground. Rocks are placed around the tripod legs for added stability.

- **Attach antenna:** A tribrach plus rotating antenna adapter is fixed to the tripod, then an optical plummet is used to make sure the tripod setup is precisely levelled and the adapter is centered above the survey mark. When the tripod setup is level the antenna is installed in place of the optical plummet, and the antenna aligned to north. The antenna cable must be connected to the receiver and then taped to the tripod leg to prevent it from blowing around during the survey.

- **Measure height of antenna:** The distance from the notch in the pin or benchmark to the outside edge of the antenna is measured using a calibrated height-stick. The height is measured at three places evenly spaced around the antenna. This 'slant height' is used in processing to calculate the direct height of the antenna above the survey mark.

- **Start receiver:** In 2004, Trimble 5700 receivers were preprogrammed by Neville Palmer (Palmer, 2004) before the survey to record at 30-second intervals, with an elevation mask of 10°. We also used Trimble 4000 and Ashtech receivers, which were programmed in the field using the same input values. The receivers are powered by one or two 12V sealed lead acid batteries. The receiver and batteries are sealed in heavy duty plastic bags to prevent them from getting wet.

- **Log sheet:** Detailed information from the station setup is recorded on a log sheet, which is later archived at GNS. These log sheets are an important source of information during data processing, in particular to help identify causes of anomalous results.

- **Data download:** Each site was surveyed for at least 24 hours. Once a site survey is completed the data are transferred from the GPS receiver to a laptop computer using Trimble software GPLOAD, or Ashtech REMOTE32.

Detailed instructions for programming Trimble and Ashtech receivers and for downloading data are found in Matheson (2001) and Palmer (2004).

Appendix B

Raukumara Peninsula GPS data processing

B.1 Raukumara Peninsula campaign GPS time series

Figure B.1 shows the locations of Raukumara Peninsula campaign GPS station that have been surveyed twice or more. Station position time series of stations surveyed more than twice are presented here (Figures B.2–B.17). These time series, showing east, north, and up components of the station position, are generated using data from the BERNESE-ADJCOORD processing, in which Auckland station AUCK was used as a base station. Uncertainties are 95% for all station positions. We use the mean station position as the origin and plotting the deviation from that point.

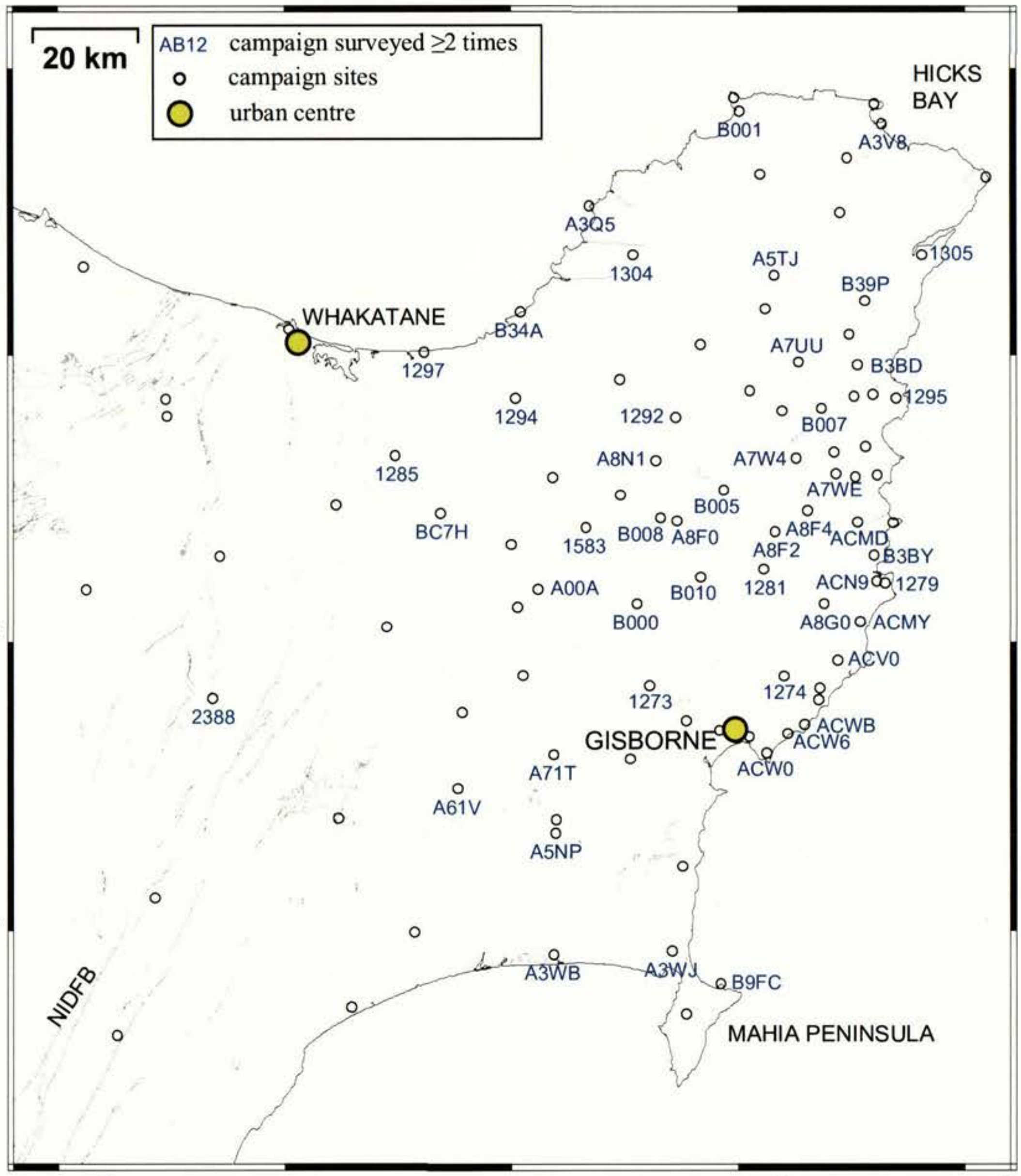

Figure B.1: **Locations of Raukumara Peninsula campaign GPS sites** that have been surveyed twice or more; sites surveyed **three times or more** are presented in this appendix as station position time series (Figures B.2–B.17).

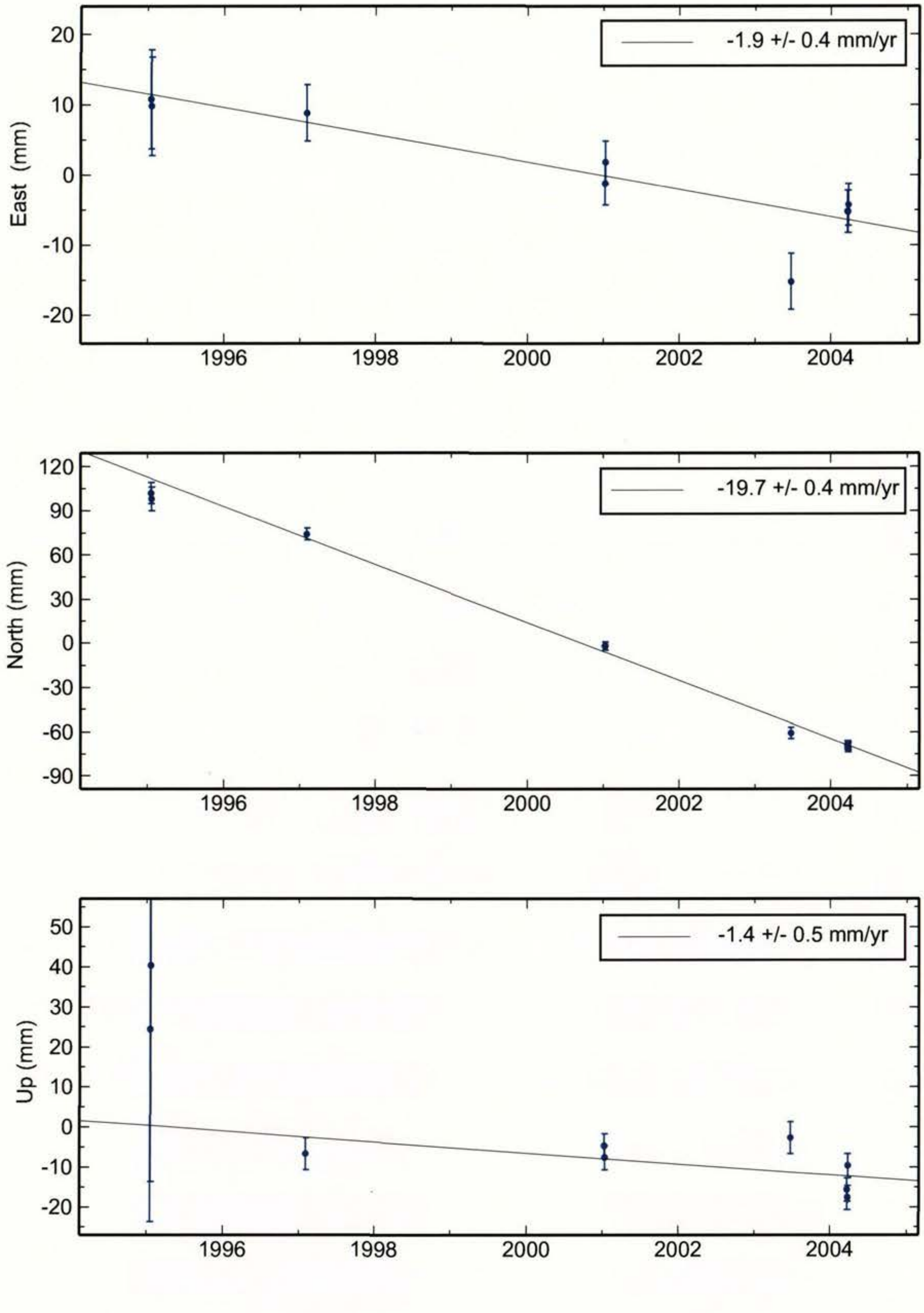

Figure B.2: Time series of station 1273

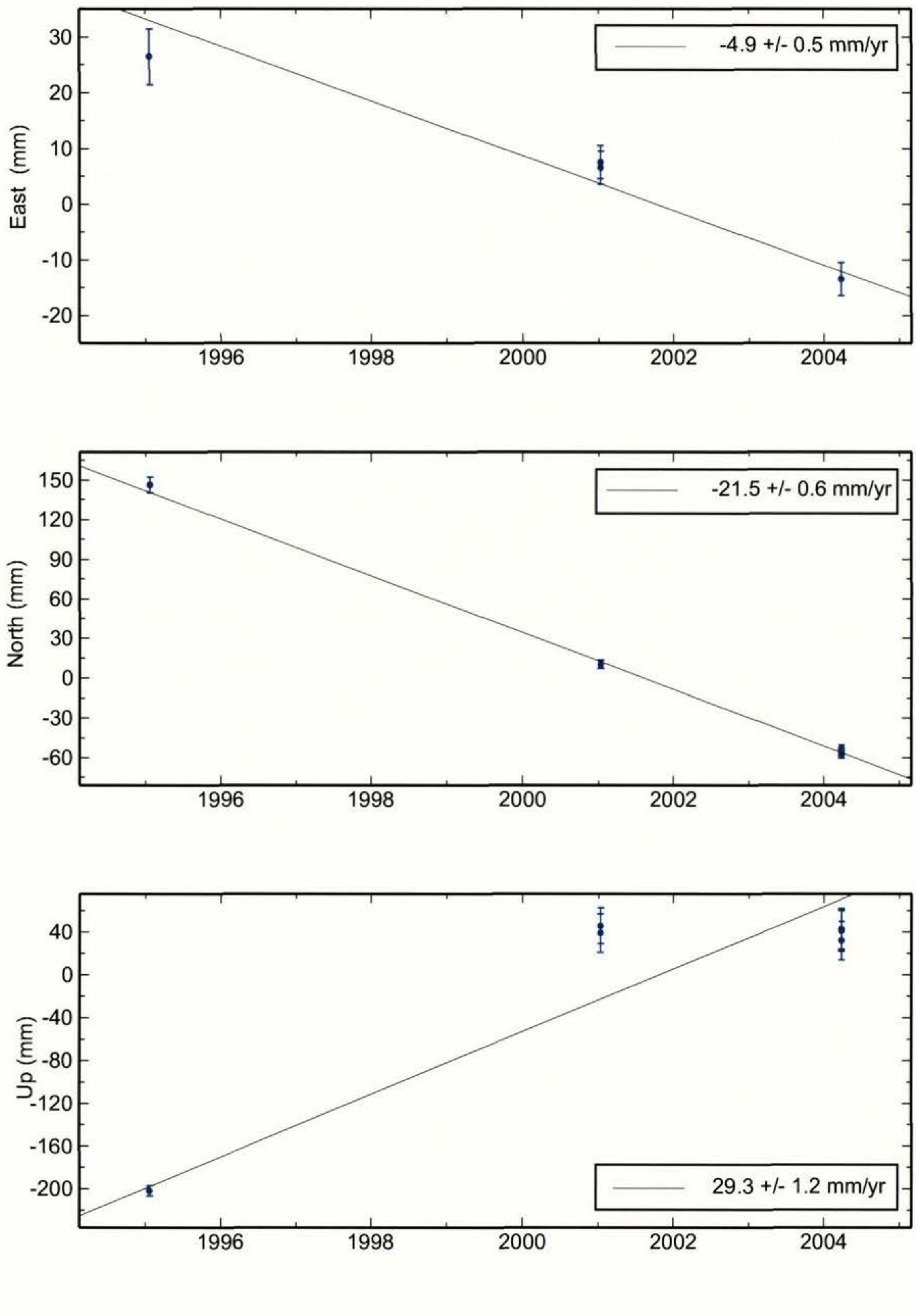

Figure B.3: Time series of station 1274

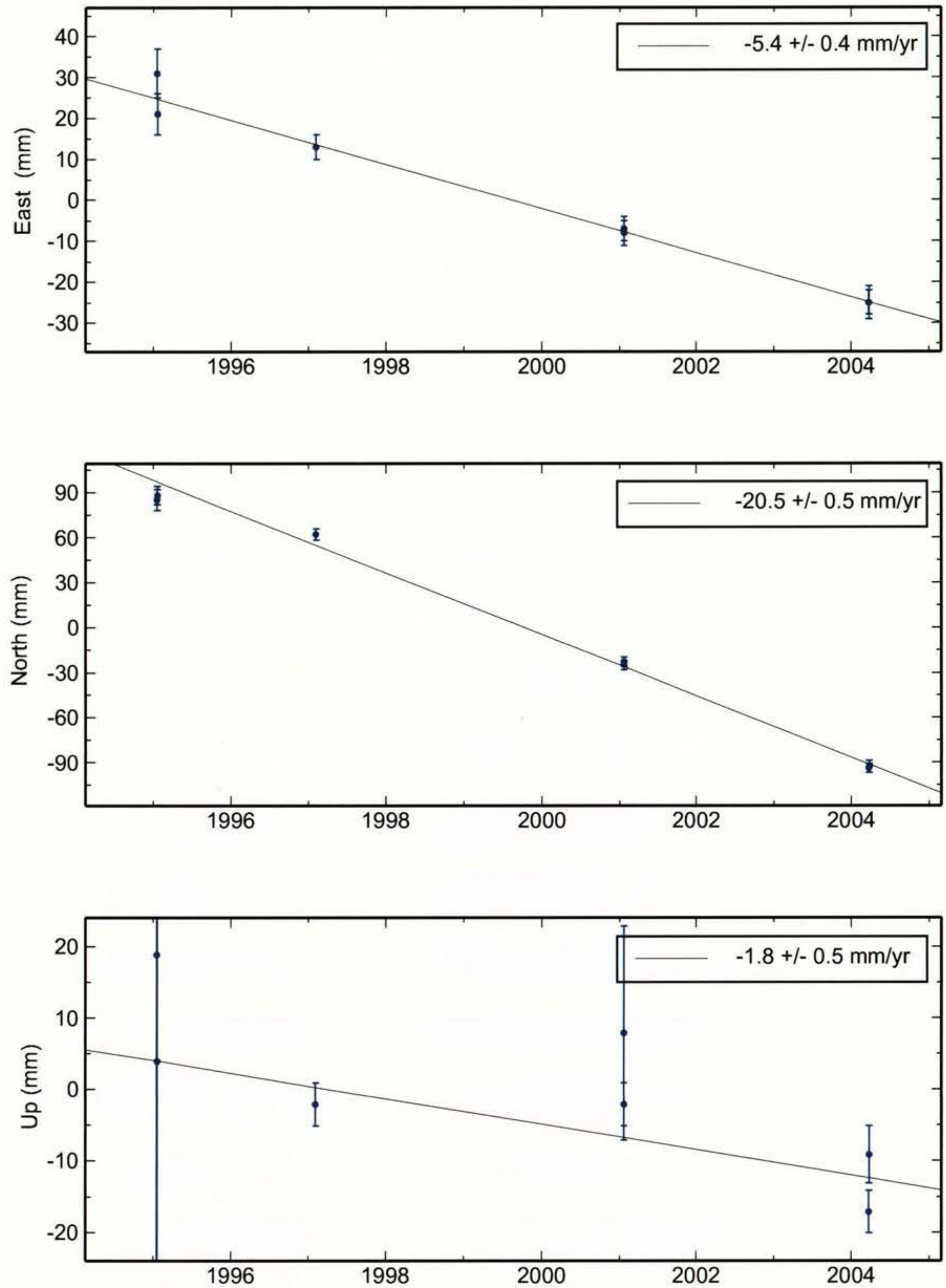

Figure B.4: Time series of station 1279

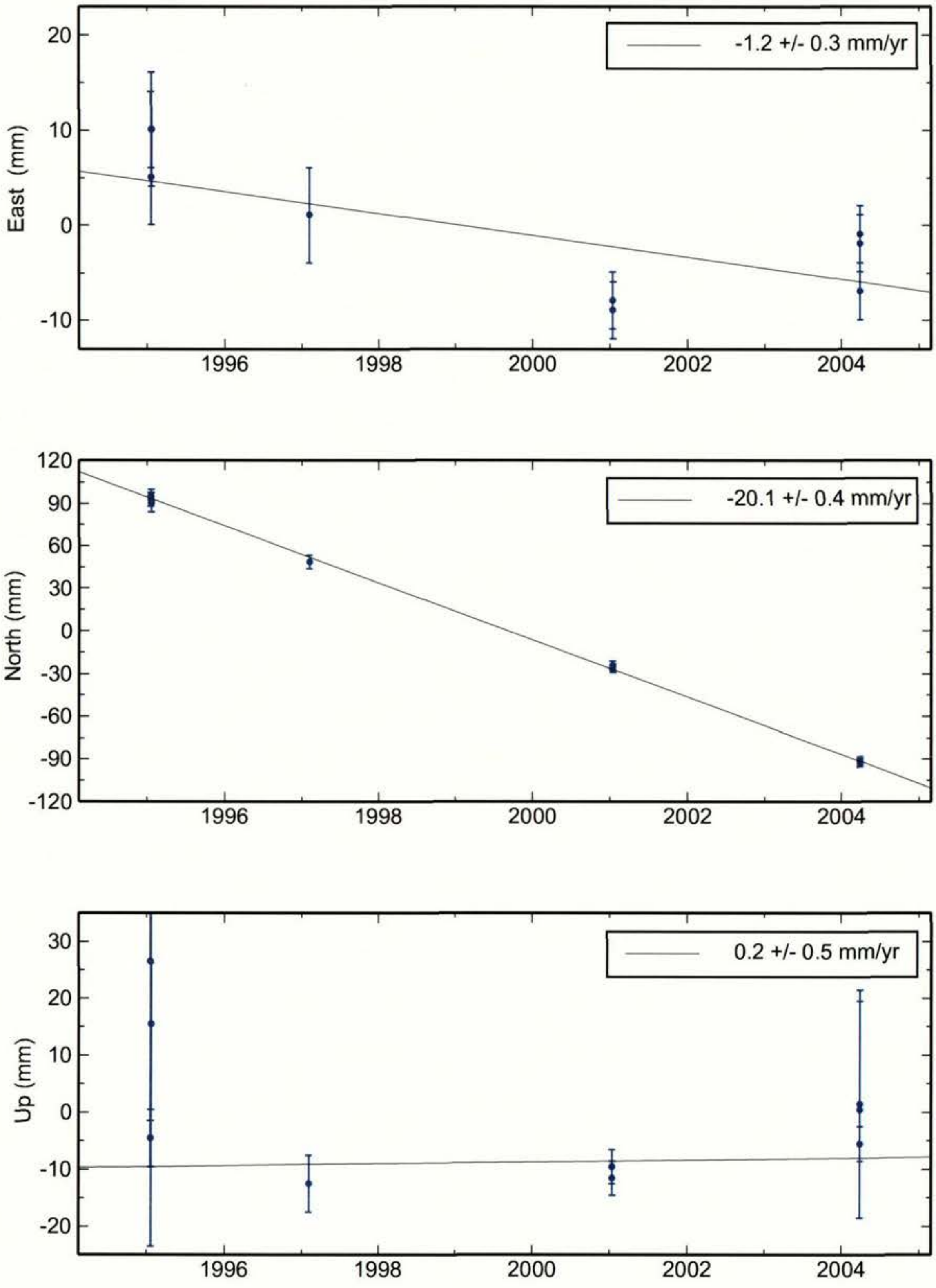

Figure B.5: Time series of station 1281

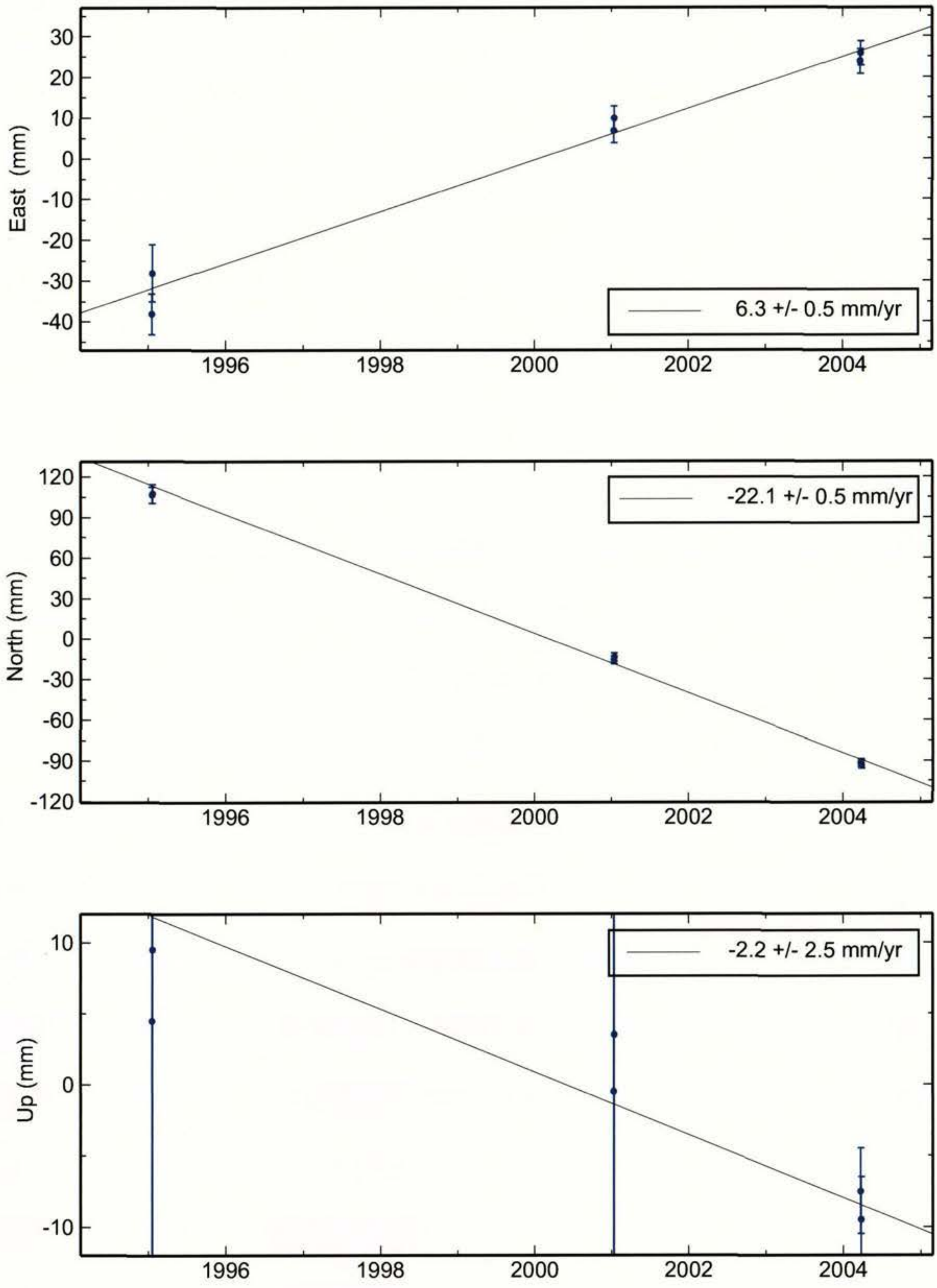

Figure B.6: Time series of station 1305

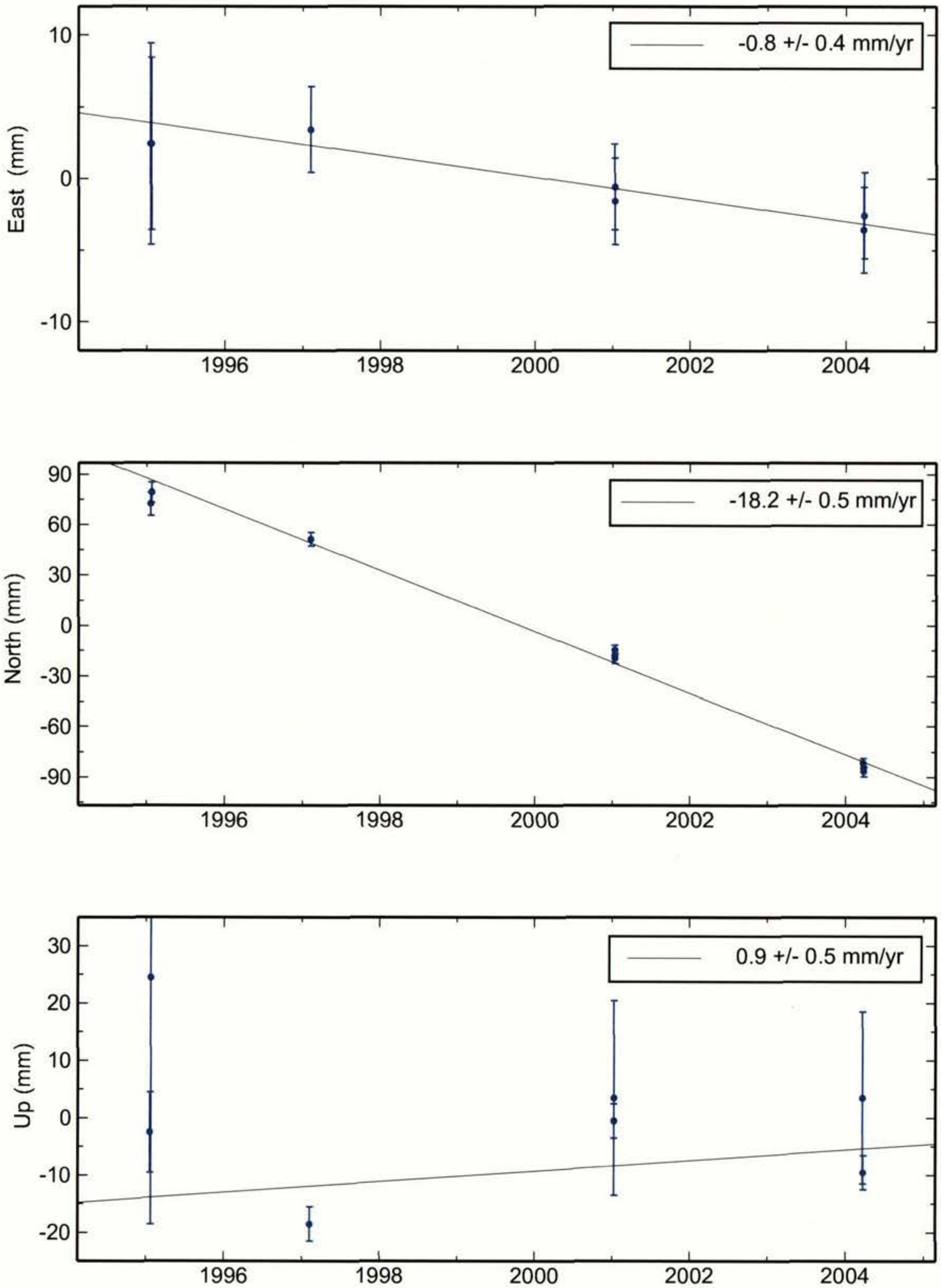

Figure B.7: Time series of station A00A

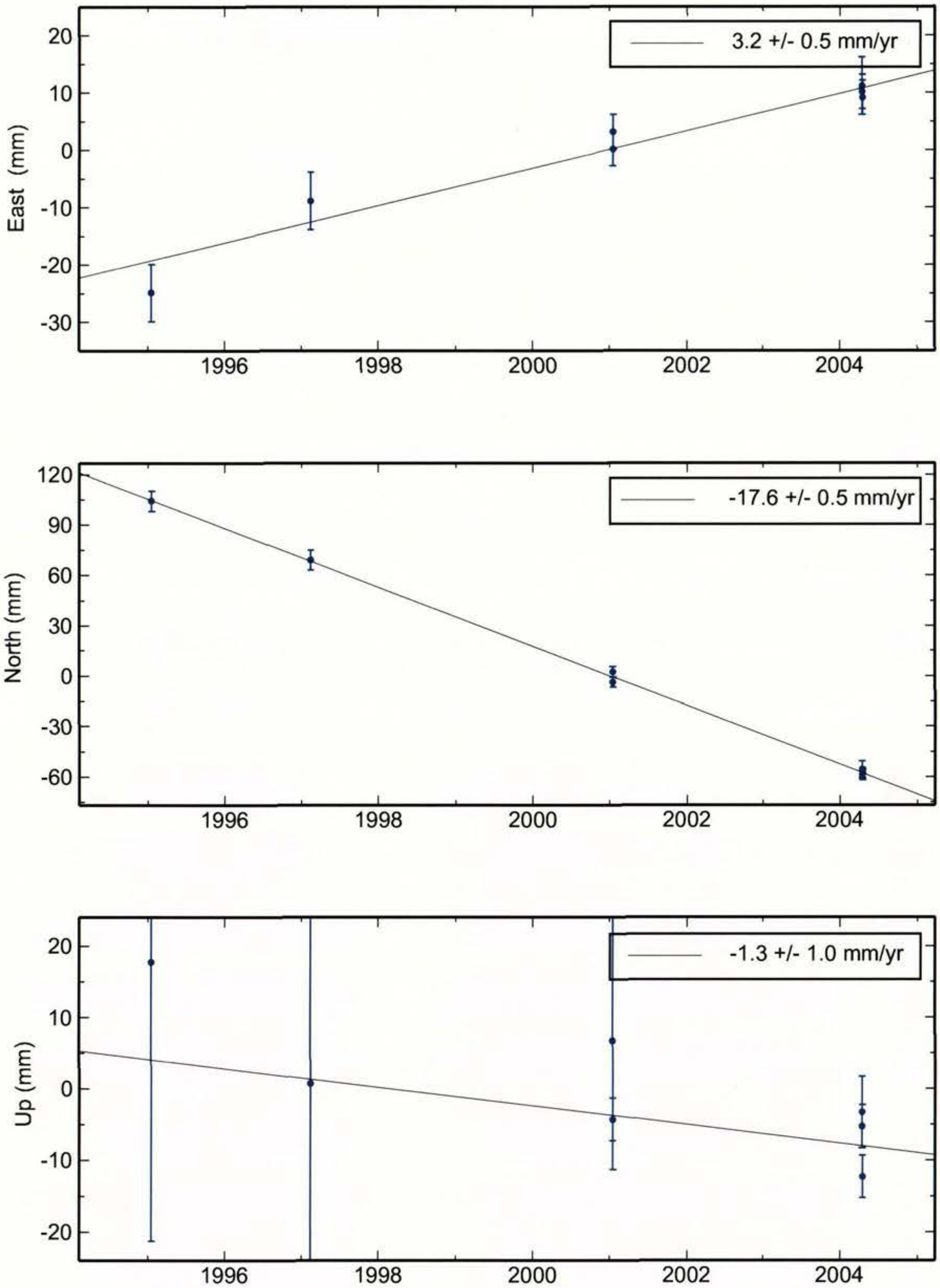

Figure B.8: Time series of station A3Q5

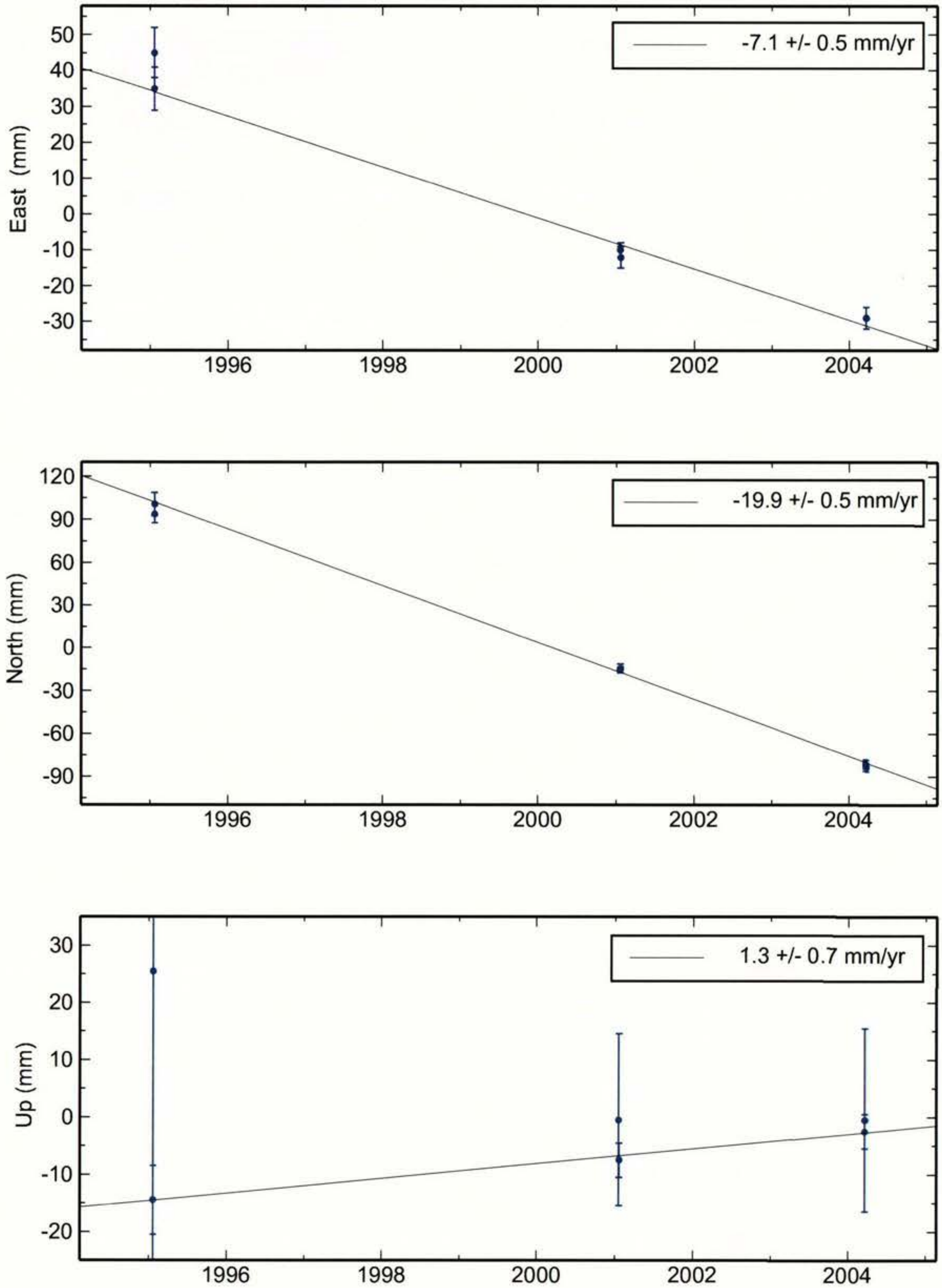

Figure B.9: Time series of station A3WJ

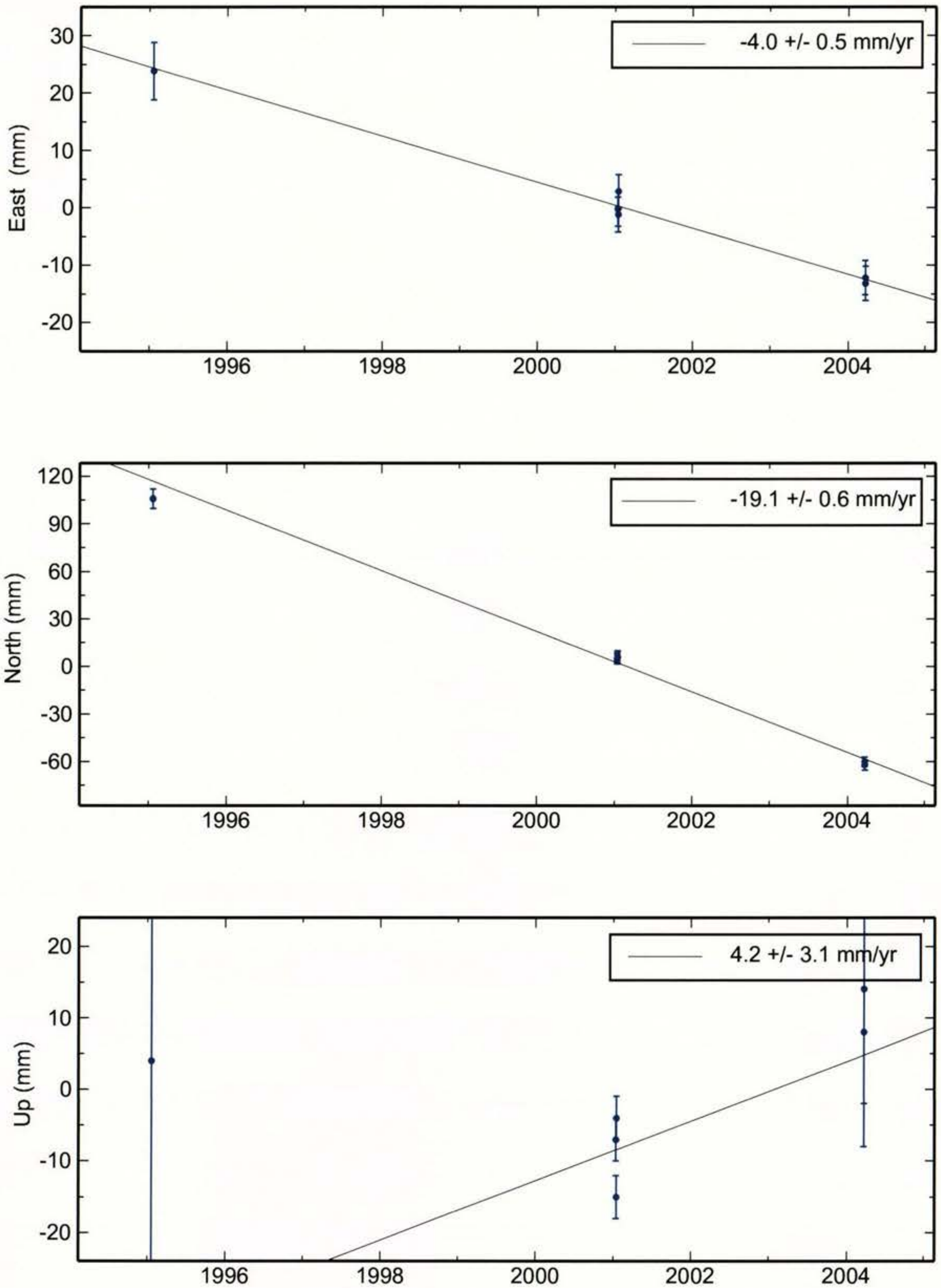

Figure B.10: Time series of station A5NP

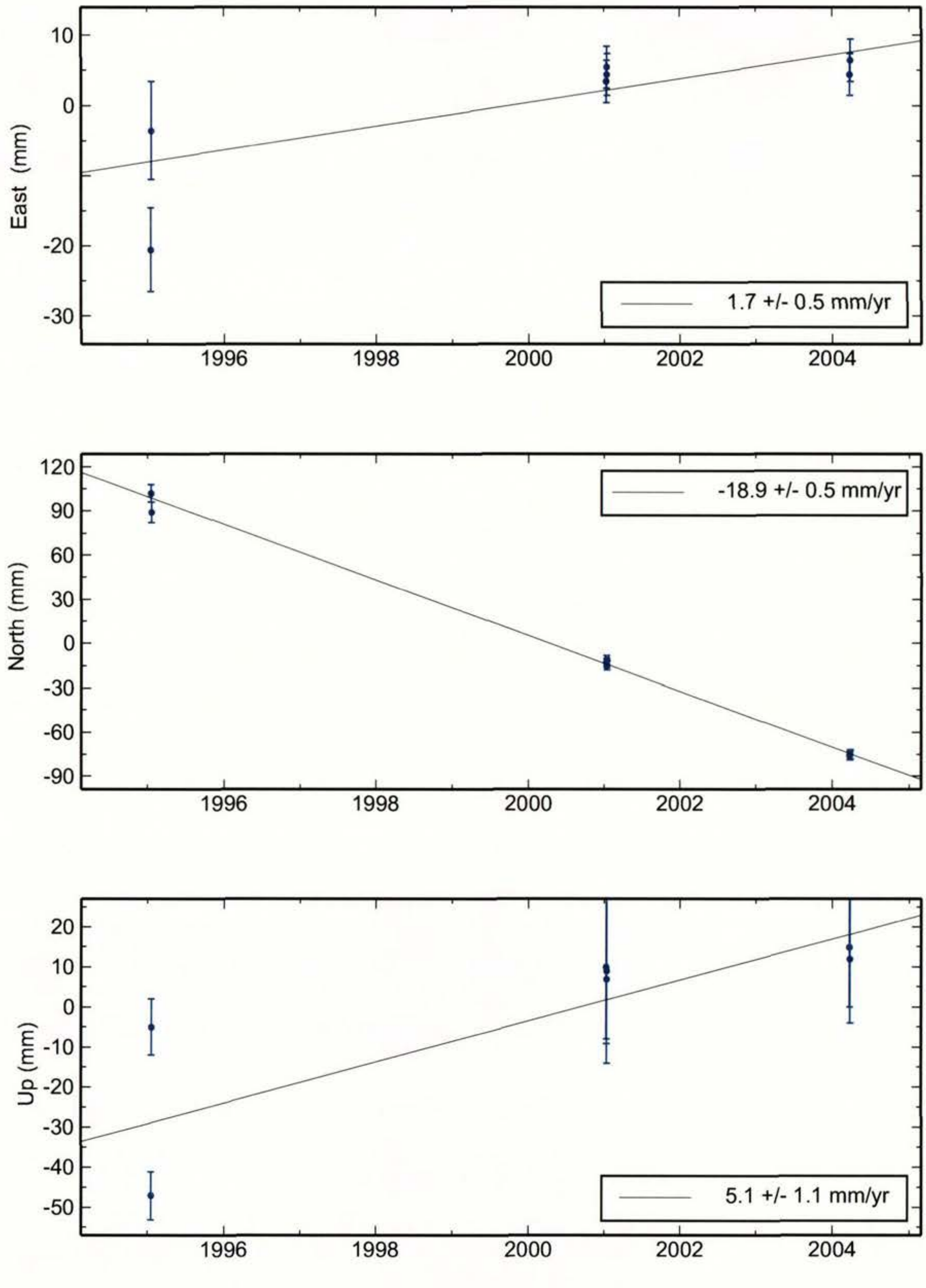

Figure B.11: Time series of station A5TJ

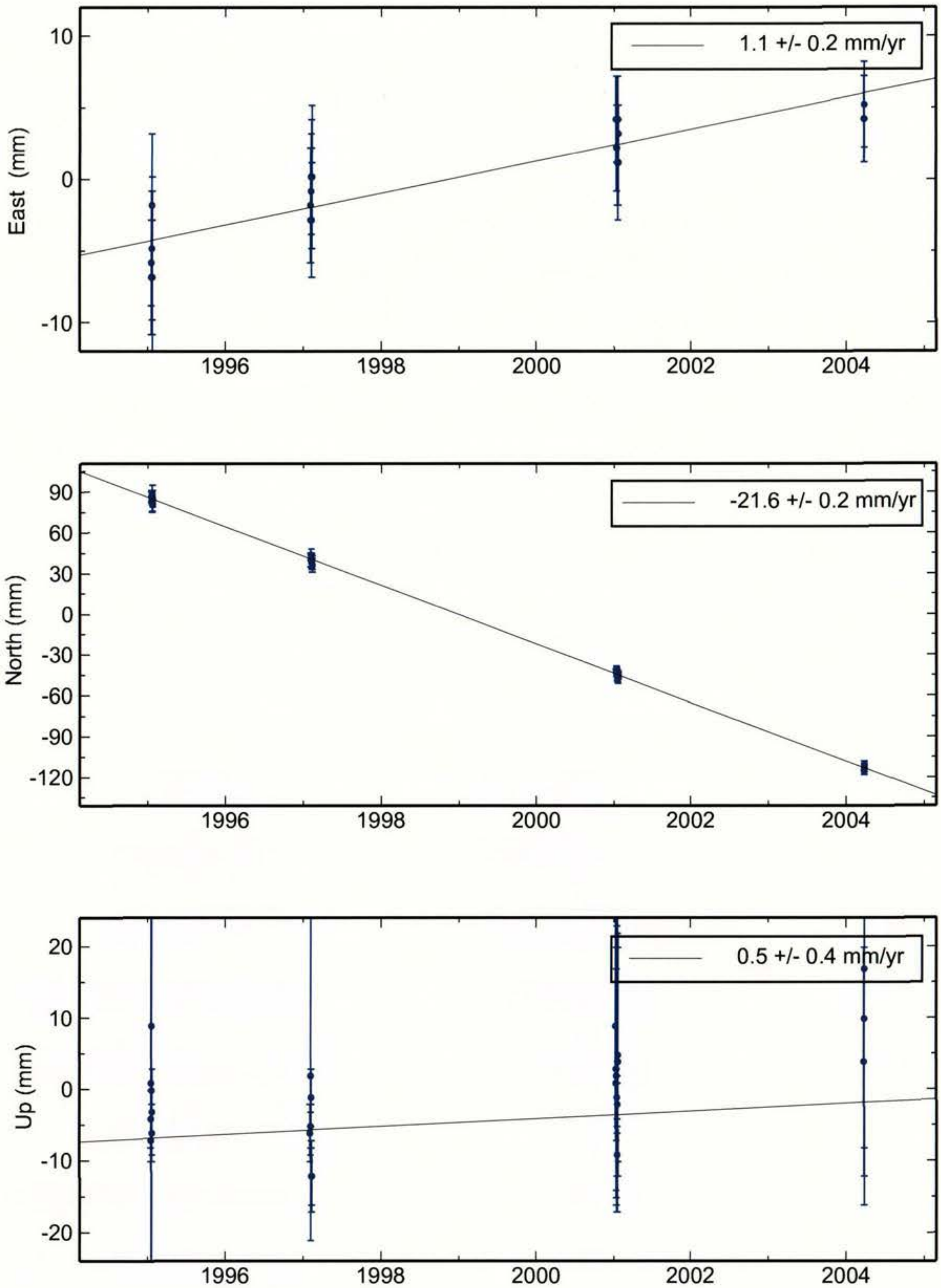

Figure B.12: Time series of station A7WE

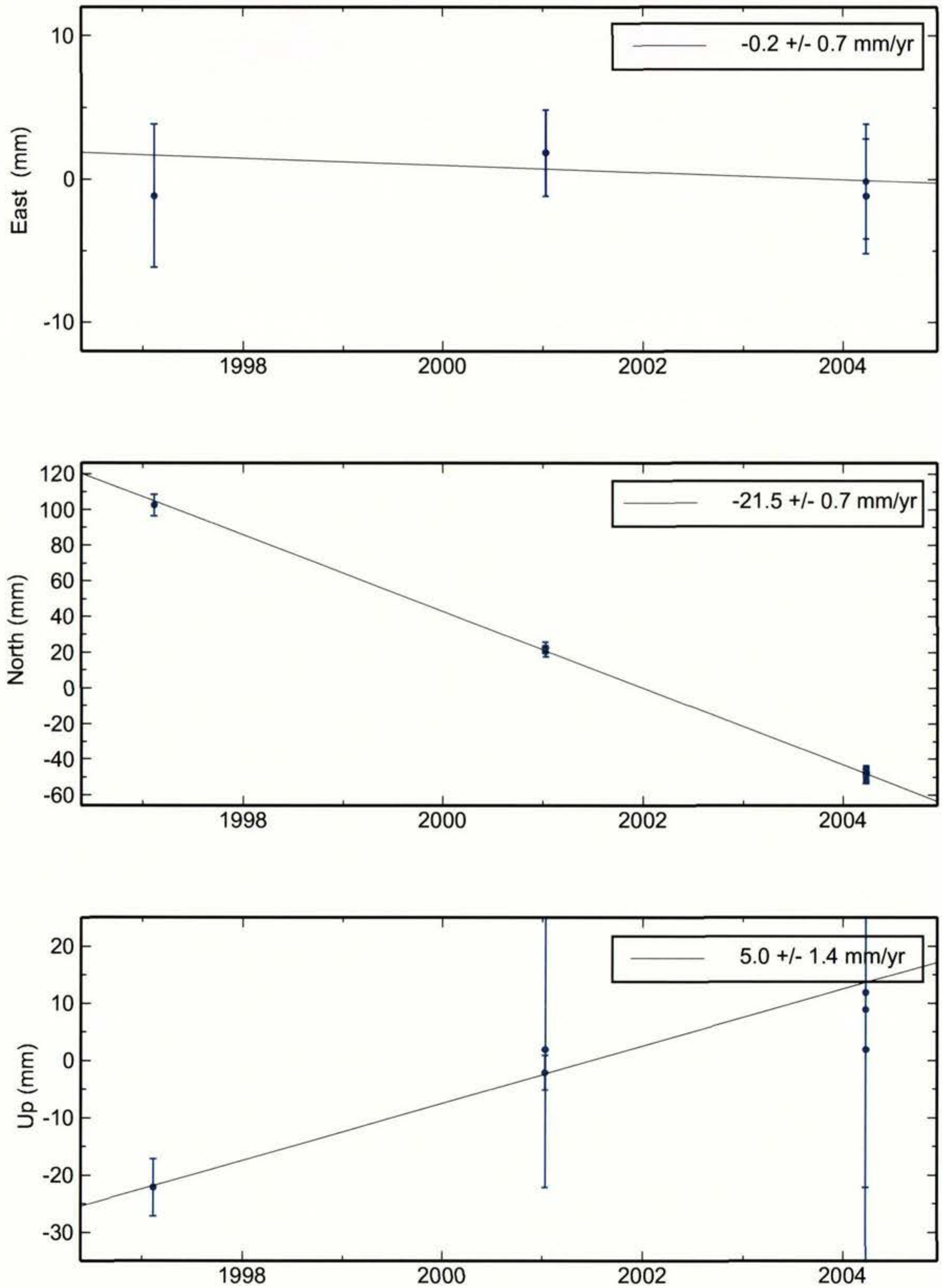

Figure B.13: Time series of station A8F4

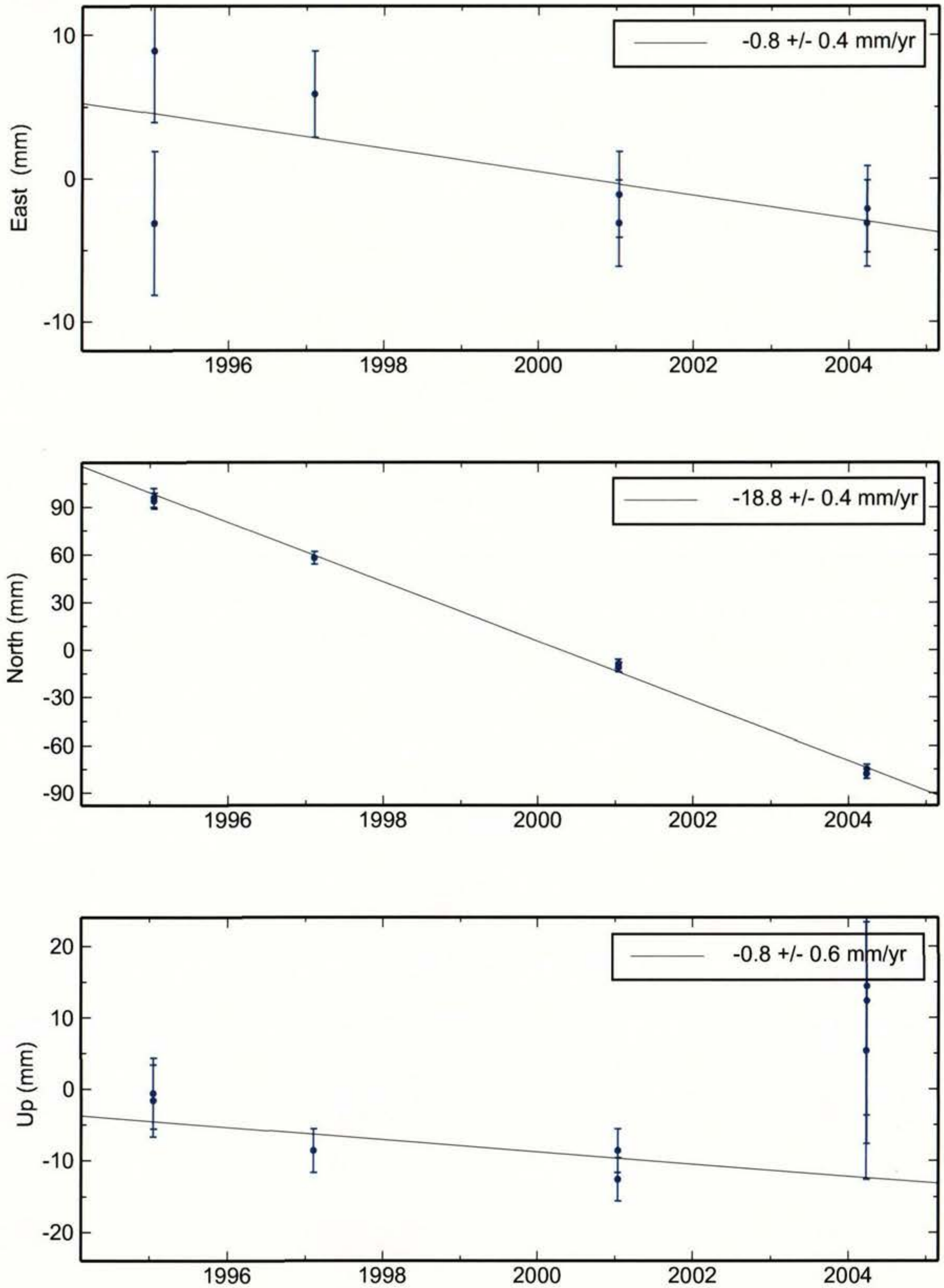

Figure B.14: Time series of station A8N1

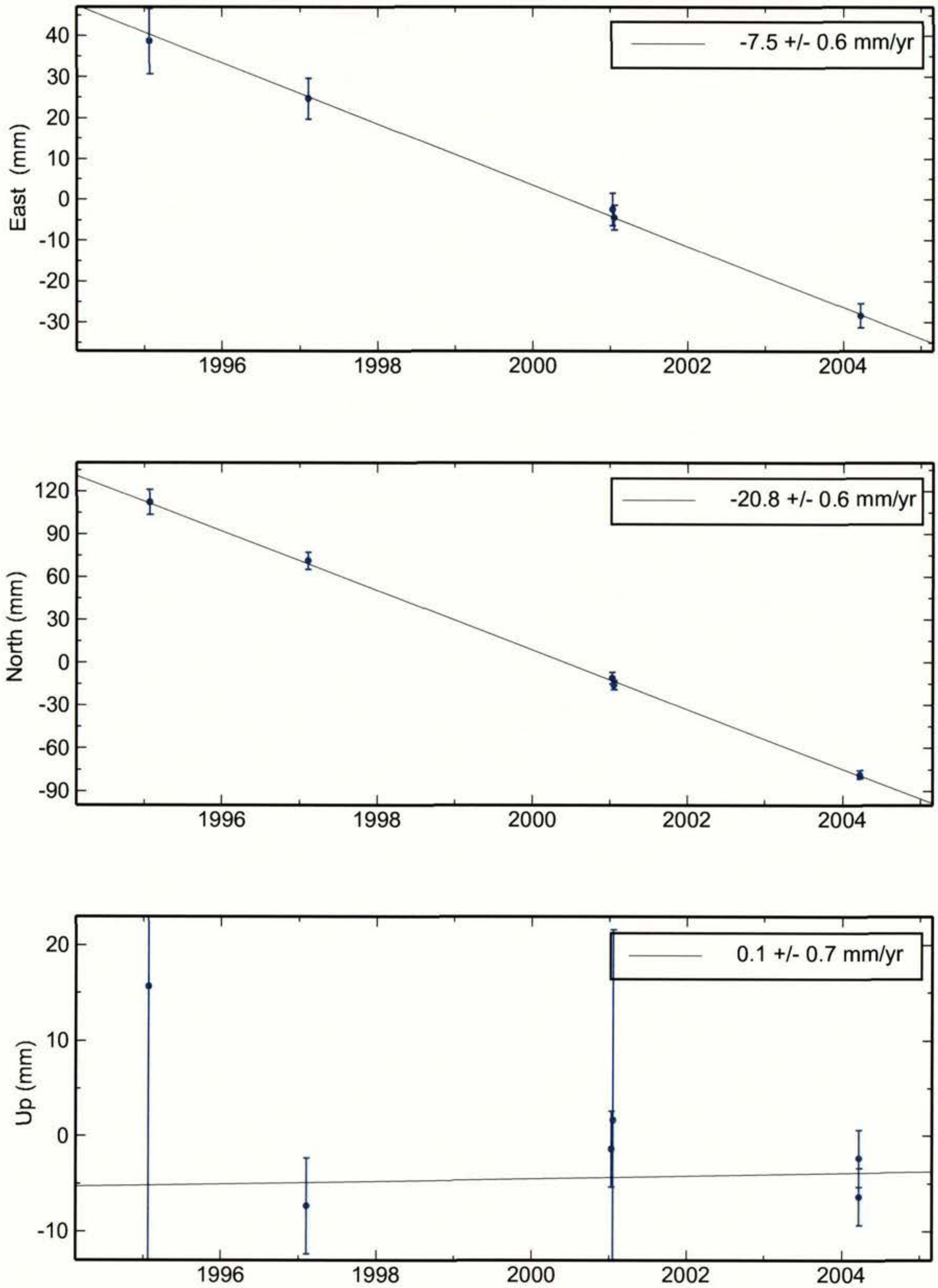

Figure B.15: Time series of station ACMY

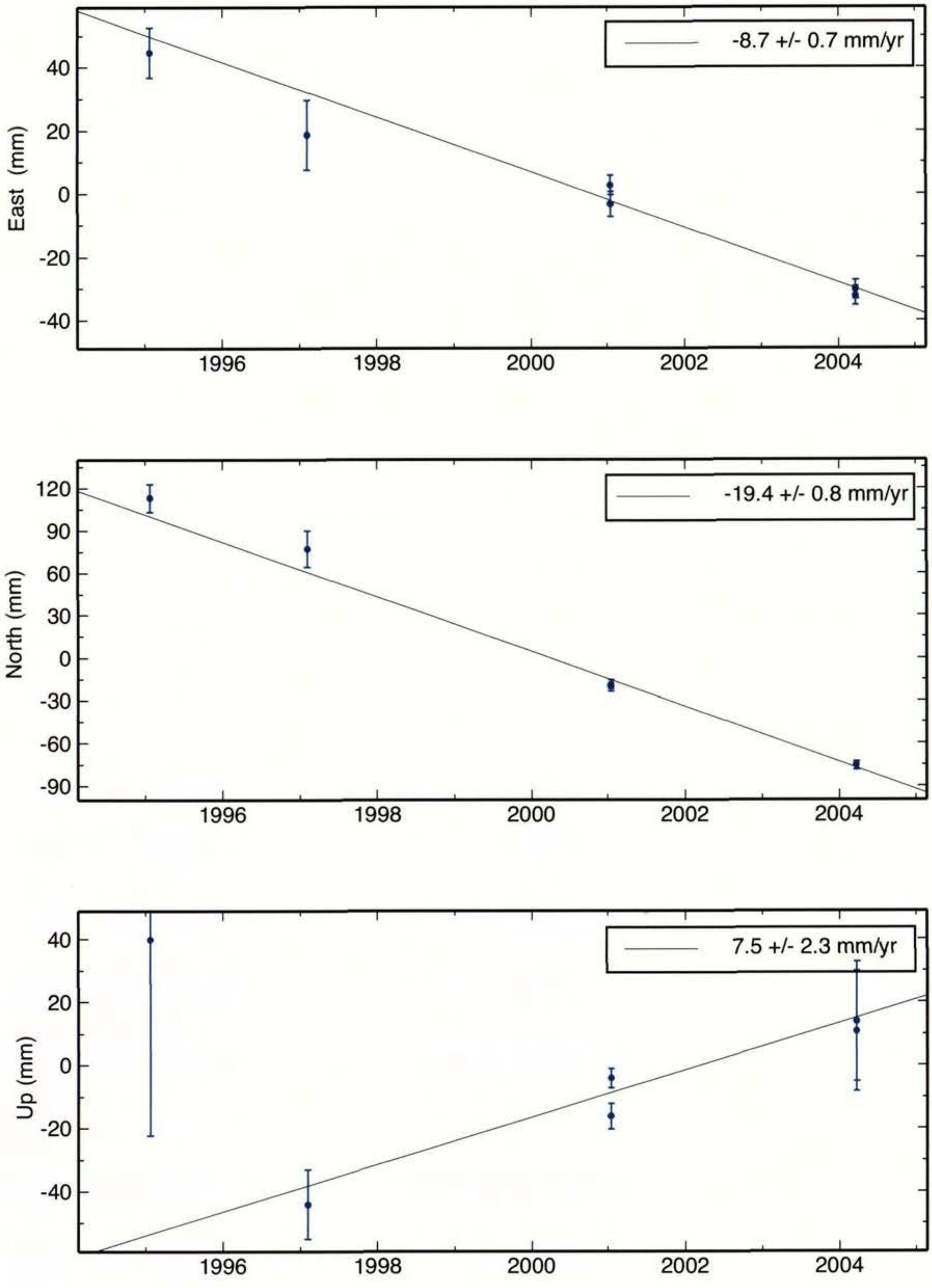

Figure B.16: Time series of station ACW0

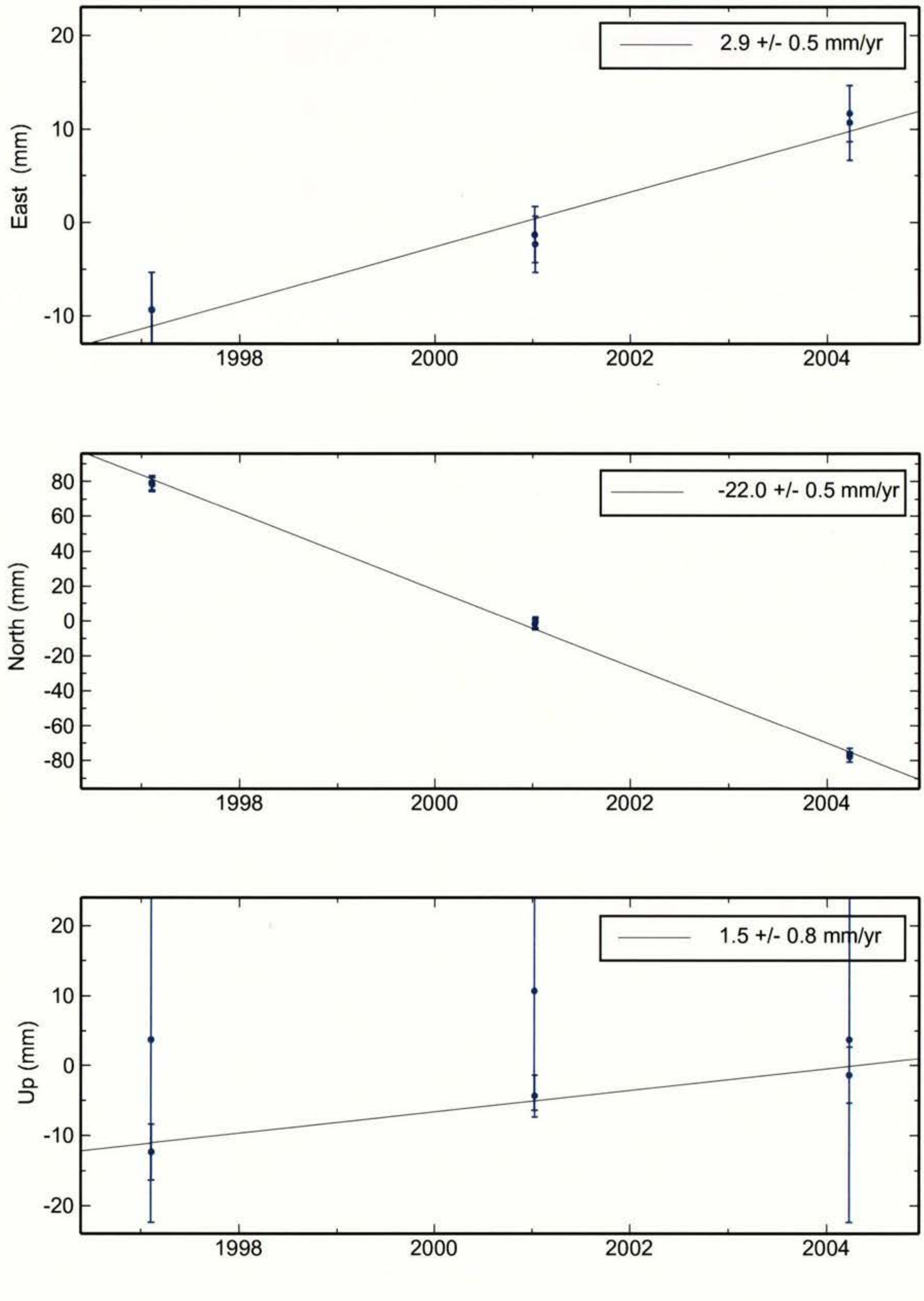

Figure B.17: Time series of station B3BD

Appendix C

Publication

GEOPHYSICAL RESEARCH LETTERS, VOL. 32, L16305, doi:10.1029/2005GL023607, 2005

Slow slip on the northern Hikurangi subduction interface, New Zealand

A. Douglas,[1] J. Beavan,[2] L. Wallace,[2] and J. Townend[1]

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[1] In October 2002, a surface displacement episode of 20–30 mm magnitude was observed over a ~10 day period on two continuous Global Positioning System (GPS) instruments near Gisborne, North Island, New Zealand. We interpret this to result from slow slip on the northern Hikurangi subduction interface. Using ten years of regional campaign GPS (1995–2004) and recent continuous GPS data, we estimate the recurrence interval for similar events to be 2–3 yrs. In November 2004, a similar slow slip event occurred within this recurrence period. The 2002 event can be modeled by ~18 cm of slow slip near the down-dip end of the seismogenic zone on the subduction interface offshore of Gisborne. The campaign GPS data show that the 2002 slow slip event had little effect on regional strain patterns. **Citation:** Douglas, A., J. Beavan, L. Wallace, and J. Townend (2005), Slow slip on the northern Hikurangi subduction interface, New Zealand, *Geophys. Res. Lett.*, *32*, L16305, doi:10.1029/2005GL023607.

1. Introduction

[2] Transient fault slip episodes, occurring over much longer time periods (days–months) than earthquakes, have been recorded with Global Positioning System (GPS) instruments at several subduction margins [e.g., *Dragert et al.*, 2001; *Obara et al.*, 2004; *Larson et al.*, 2004]. These so-called slow slip events may make a significant contribution to moment release in subduction zones; quantifying their size is therefore a key task in characterizing seismic hazard at subduction zones.

[3] The Raukumara Peninsula, the easternmost part of New Zealand's North Island, is an important location for observations of the northern Hikurangi subduction margin (Figures 1a and 1b). Offshore to the east, the Hikurangi Trough marks the present-day plate boundary where the Pacific plate subducts beneath the North Island. The majority of earthquakes beneath the peninsula occur in a westward-dipping zone defining the down-going Pacific plate [*Ansell and Bannister*, 1996]. To the west, back-arc extension forms the Taupo Volcanic Zone (TVZ) [e.g., *Wilson et al.*, 1995], separated from the Raukumara Peninsula by the North Island Dextral Fault Belt (NIDFB) (Figure 1b) [*Beanland and Haines*, 1998].

[1]School of Earth Sciences, Victoria University of Wellington, Wellington, New Zealand.

[2]Institute of Geological and Nuclear Sciences, Lower Hutt, New Zealand.

0094-8276/05/2005GL023607$05.00

[4] In October 2002, a rapid (compared to normal plate motion) surface deformation event of 20–30 mm magnitude was observed over an ~10 day period on two continuous GPS (CGPS) instruments near Gisborne (GIS1 and GISB; Figure 1c). The event, interpreted as the result of slow slip on the Hikurangi subduction interface by *Beavan et al.* [2003], occurred during the early stages of CGPS network development in the region. The spatially limited CGPS observations do not constrain the margin-parallel extent of this event. Campaign GPS data from the Raukumara Peninsula may allow better spatial control of the 2002 event and detection of earlier deformation episodes, if the spatial and temporal sampling of the campaign GPS are sufficient [e.g., *Larson et al.*, 2004].

2. Data

2.1. Campaign GPS Data Analysis

[5] We analyzed GPS data from surveys conducted on the Raukumara Peninsula in January 1995 [*Árnadóttir et al.*, 1999], February 1997 and January 2001, plus a new campaign we conducted in March 2004. To ensure consistency between the results, we used identical processing techniques for all data sets.

[6] In each campaign, data lengths of 24 hours or more were obtained at most sites. The GPS phase data were processed using standard methods and Bernese 4.2 software [*Beutler et al.*, 2001] to determine daily estimates of relative coordinates and their covariance matrices. International GPS Service (IGS) final ITRF2000 orbits and associated polar motion files were held fixed and one station's coordinates (AUCK; Figure 1a) were tightly constrained during each day's processing. AUCK was not installed at the time of the 1995 campaign so we used a tie from a nearby (375 m) site. All daily coordinate-difference solutions and their covariances were input into least squares software ADJCOORD [*Crook*, 1992] to check for outliers and construct station position time series. At this stage, stations with very short observing sessions (some 1995 sessions were as short as 4–5 hours), or with site velocities grossly inconsistent with neighboring sites, were removed from the solutions (less than 5% of sites).

2.2. GPS Time Series

[7] Figure 2a shows station position time series relative to AUCK on the stable Australian plate. The east component of the GISB time series from 2002 to 2005 (Figure 2b) reveals three rapid surface displacement episodes; the first and third were similar in size (~20–30 mm), while the second was a factor of five smaller and of longer duration. The slow slip event displacements are in an ESE direction, opposite to the WNW direction of subduction of the Pacific

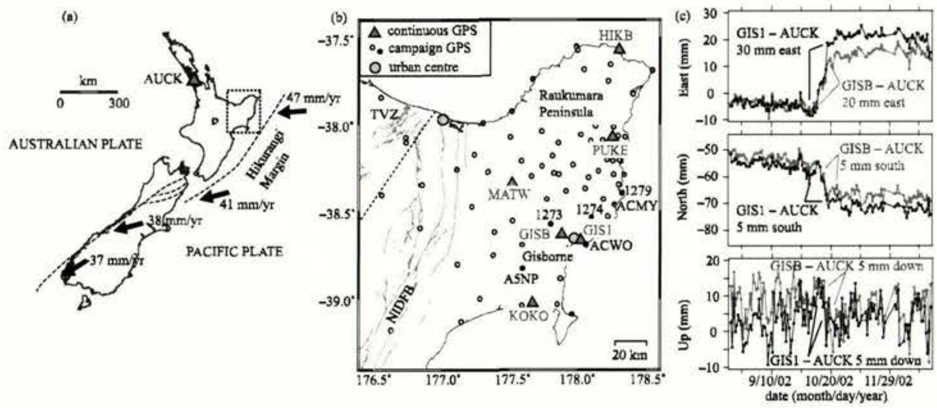

Figure 1. (a) Map showing study area and Pacific–Australia relative velocity vectors [*DeMets et al.*, 1994]. (b) Current Raukumara Peninsula continuous and campaign GPS stations. TVZ — Taupo Volcanic Zone, NIDFB — North Island Dextral Fault Belt. Time series from six campaign sites (solid circles) are shown in Figure 2a. The 2002 slow slip event (c) was recorded on CGPS stations GIS1 and GISB, installed in August 2001 and July 2002 respectively; the other CGPS stations were installed after 2003.

plate beneath the Raukumara Peninsula [*Wallace et al.*, 2004].

3. Results

3.1. Constraining the 2002 Event With Campaign GPS

[8] We find that the several-year time intervals between GPS campaigns on the Raukumara Peninsula make it impossible to identify transient motion from the 2002 slow slip event, or any previous slow slip events, in the campaign time series (Figure 2a). The shortest interval between GPS campaign measurements has been 2 years, while it took approximately 1.5 years for the eastward displacement caused by the 2002 event at GISB to be recovered by westward motion at that site. We therefore cannot use the campaign time series to help constrain the spatial extent or moment release of the 2002 slow slip event. However, we can use the campaign data to estimate slow slip recurrence intervals.

3.2. Slow Slip Recurrence Intervals

[9] Quasi-periodic recurrence intervals have been observed for slow slip events in the Cascadia subduction margin [e.g., *Rogers and Dragert*, 2003; *Szeliga et al.*, 2004], and the Guerrero region of Mexico [*Lowry et al.*, 2005]. Recurrence intervals have been used successfully in Cascadia to predict the timing of future events.

[10] To estimate a recurrence interval for slow slip events similar to the Gisborne 2002 event, we compare the 'long-term' rates averaged over the ten year span of the campaign time series, with the short-term rates between slip events. This short-term rate is estimated by averaging the best fit linear trends of the CGPS time series during times when no slow slip is taking place. We choose pairs of neighboring stations ACW0 (campaign) and GIS1 ($\sim$5 km apart), and 1273 (campaign) and GISB ($\sim$10 km apart), which are likely to have similar displacement histories (Figure 2a). As stations 1273 and GISB are further apart we have less confidence in this comparison.

[11] We estimate the average rate at GIS1 between the 2002 and 2004 events to be 18.3 ± 1.9 mm/yr west relative to AUCK. Based on a long-term rate (1995–2004) at ACW0 of 8.7 ± 0.7 mm/yr west, the deficit in westward motion at ACW0 is 9.6 ± 2.0 mm/yr. We assume the deficit is due to episodic slow slip events. The 2002 event resulted in $\sim$30 mm eastward displacement at GIS1, implying a recurrence interval of $\sim$3.1 ± 0.6 yrs, assuming the 2002 event has typical surface displacement. Comparing stations 1273 and GISB, we find a deficit in westward motion of 10.9 ± 1.2 mm/yr. The 2002 event resulted in $\sim$20 mm eastward displacement at GISB, giving a recurrence interval

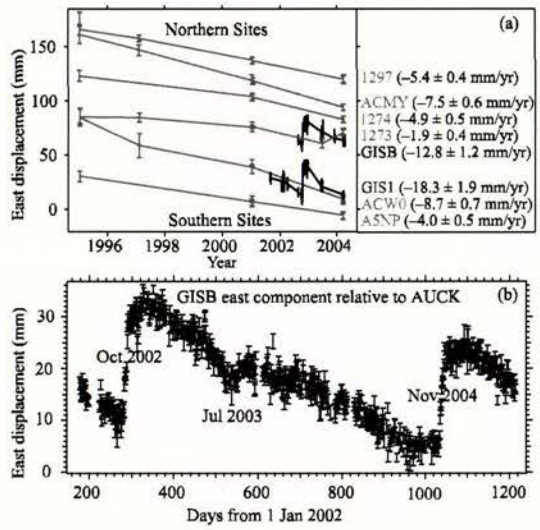

Figure 2. (a) Position time series and velocities relative to AUCK averaged over 1995–2004 interval for selected campaign sites (Figure 1b), and averaged over selected days for CGPS stations GIS1 and GISB. Rates for GIS1 and GISB are estimated by averaging the best fit linear trends during periods when no slow slip is taking place. (b) GISB time series from 2002–2005 showing eastward surface displacements of 20–25 mm in October 2002 and November 2004, and 5 mm in July 2003. See color version of this figure in the HTML.

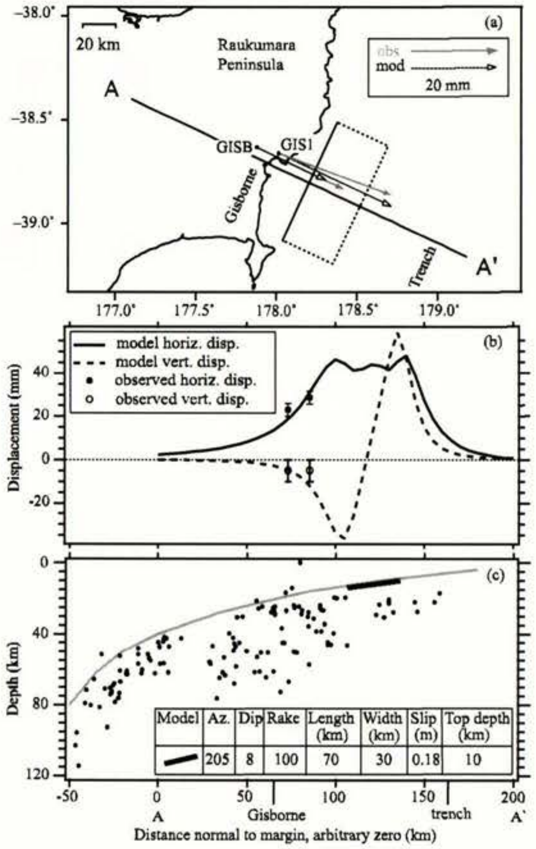

Figure 3. (a) Model slip plane and observed and predicted displacements for 2002 event. Dashed edges of slip plane are not well constrained. (b) Observed and predicted displacements along A–A′. (c) Profile along A–A′ showing model slip plane, current seismicity and subduction interface.

of ~1.8 ± 0.2 yrs. The uncertainties in these recurrence estimates are probably underestimated as we have not incorporated the uncertainty in magnitude of the slow slip events.

[12] Our estimates point to slow slip events similar to the October 2002 event occurring every 2–3 years. In November 2004, an event (Figure 2b) of similar surface displacement was recorded. The timing of this event, ~2.1 years after the previous one, is consistent with our predicted recurrence interval. Smaller events, such as that in July 2003 (Figure 2b), may modulate the recurrence interval. Though three more continuous GPS stations had been installed in the Raukumara region by November 2004, they were either too far away or too recently installed to provide data to constrain that event.

3.3. Forward Modeling of the Gisborne 2002 Slow Slip Event

[13] We interpret the observed displacements for the 2002 slow slip event at GIS1 and GISB using an elastic dislocation model comprising uniform slip on a plane in an elastic half space (Figure 3). The campaign displacements do not provide useful data for this modeling, as discussed above.

[14] We model slip on different parts of the subduction interface by varying the dip, depth and position of the slip plane in accord with the interface geometry inferred from seismology [*Reyners*, 1998] (Figure 3c). We vary the rake, width and slip by trial and error for each different slip plane location.

[15] One plausible model requires ~18 cm of thrust slip on a plane whose lower edge is at depth ~14 km on the subduction interface ~20 km east of Gisborne (Figure 3a). This model fits the observed horizontal data quite well (Figure 3b) and the downward displacement of the Gisborne CGPS sites. The slip magnitude and lower edge of the event are reasonably well constrained, but the poor spatial distribution of GPS data prevents us constraining its up-dip or lateral extent. We rule out slip on the deeper subduction interface (down-dip of GISB) as those models predict uplift for the Gisborne region.

[16] While fault planes smaller than that in Figure 3 can fit the data equally well, the larger slip required, combined with our estimated slow slip recurrence interval of 2–3 yrs, implies that the rate of slip in slow events would exceed the long-term rate of convergence between the Pacific plate and the forearc block of ~54 mm/yr up-dip [*Wallace et al.*, 2004].

[17] The slip region in the model of Figure 3 is near the downdip end of the seismogenic zone on the subduction interface [*Wallace et al.*, 2004]. This is similar to the location of slow slip events observed in SW Japan and Cascadia [e.g., *Dragert et al.*, 2001; *Obara et al.*, 2004], though the down-dip end of inter-seismic coupling at the north-eastern Hikurangi subduction zone is comparatively shallow [*Reyners*, 1998; *Wallace et al.*, 2004].

3.4. Temporal Variation of Strain

[18] *Árnadóttir et al.* [1999] studied temporal variations in maximum shear strain rates on the Raukumara Peninsula using triangulation data from 1925 and 1976 and campaign

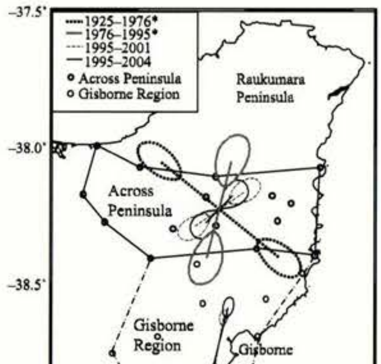

Figure 4. Temporal variation in maximum shear strain rate and azimuth of principal axis of relative extension across the Raukumara Peninsula (from triangulation and GPS between 1925 and 2004;* data from *Árnadóttir et al.* [1999]), and for the Gisborne region (from 1995–2004 GPS). Uncertainties are 2-D 68% confidence regions taking account of random error and the misfit to a uniform strain-rate model.

GPS from 1995. They interpreted variations in maximum shear strain rate and orientation between the 1925–1976 and 1976–95 epochs to result from aseismic slip over the entire shallow subduction interface to 30 km depth between the 1920s and 1970s. To investigate strain rates since 1995, we calculate the magnitude of the maximum shear strain rate and the orientation of the principal extension axis over the Raukumara Peninsula using all available campaign GPS data (Figure 4). Our strain calculation regions are based on those of *Árnadóttir et al.* [1999].

[19] Figure 4 shows there was no significant change in shear strain rate either across the Raukumara Peninsula or in the Gisborne region between 2001 and 2004, over a period when slow slip events are known to displace CGPS sites in the Gisborne region. For the 'Across Peninsula' region, there is a small but not statistically significant clockwise rotation of the 1995–2001 and 1995–2004 strain-rate azimuths as compared to 1976–1995. However, these three directions all differ significantly from the 1925–1976 direction (Figure 4).

[20] The slip on the subduction interface modeled by *Árnadóttir et al.* [1999] to explain temporal variation in strain between 1925–1976 and 1976–1995 was of wide lateral extent and reached to 30 km depth. It is very different in character to the Gisborne slow slip events recorded in 2002–2004. In our model for the 2002 event, the subduction interface slips to a depth of only $\sim$15 km, and the slipping region may be localised near the bottom of the seismogenic zone.

4. Conclusions

[21] Low temporal sampling in station position time series from the Raukumara Peninsula campaign GPS data set aliases the Gisborne 2002 slow slip event, seen as 20–30 mm of surface displacement at CGPS sites in the region. While we cannot use the campaign GPS data to constrain the spatial extent of the 2002 event, we can estimate that such events may recur at 2–3 year intervals, compatible with the timing of the November 2004 event.

[22] Elastic dislocation models show the 2002 event to be consistent with slip on the subduction interface offshore of Gisborne. We cannot accurately constrain the north–south or up-dip extents of the modeled slip plane, but we find that the down-dip extent is consistent with slip occurring near the down-dip limit of the seismogenic zone.

[23] We find that the 2002 event did not have a significant effect on regional strain-rate patterns over the Raukumara Peninsula.

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J. Beavan and L. Wallace, Institute of Geological and Nuclear Sciences, Lower Hutt, New Zealand.

A. Douglas and J. Townend, School of Earth Sciences, Victoria University of Wellington, Wellington 6001, New Zealand. (annie.douglas@gmail.com)