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Seismic Activity in the Transantarctic Mountains - Results from a Broadband Array Deployment

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Abstract - Seismic events have been recorded in the Transantarctic Mountains using a temporary 10-station broadband seismometer array. About 50% of a total of 160 events were located to the north of the array, in the vicinity of the David Glacier, inland from the Drygalski Ice Tongue, while 40% were located near the transition between the Mulock glacier and the Ross Ice Shelf, just south of the Skelton glacier. The northern events are roughly aligned in a ESE-WNW direction, sub-parallel to faults which are postulated to lie near to the events, and close to the postulated David glacier lineament. The events may be associated with deep-seated faulting along the lineament, although alternative source mechanisms involving ice movement are also possible.

INTRODUCTION

The Transantarctic mountain range (TAM, Fig.1) is an impressive morphological feature which marks the border of the East Antarctic craton and the western Ross Sea region, reaching elevations of over 4000 metres and extending along the entire length of the West Antarctic Rift System (Wilson & Finn, 1996). Most current models postulate that the TAM represent a rift shoulder which formed in response to rift-related processes in the West Antarctic Rift System. Basement rocks in the TAM consist of early Paleozoic to a late PreCambrian metamorphic-plutonic complex. Overlying the basement is a sub-horizontal layered sequence of Devonian-Jurassic sedimentary rocks ranging up to 3500m in thickness.

To the east and immediately adjacent to the TAM is the Victoria Land Basin (VLB, Fig.1), which was originally defined by marine seismic reflection data. Cooper et al. (1987) interpreted up to 14 km of stratified rock with block-faulted basement structure, typical of rift basins, and a thickness of at least 4-5 km of Cenozoic sediments. The VLB is inferred to be underlain by rifted continental crust, based on the presence of normal faulting (Cooper et al.,1987), thin

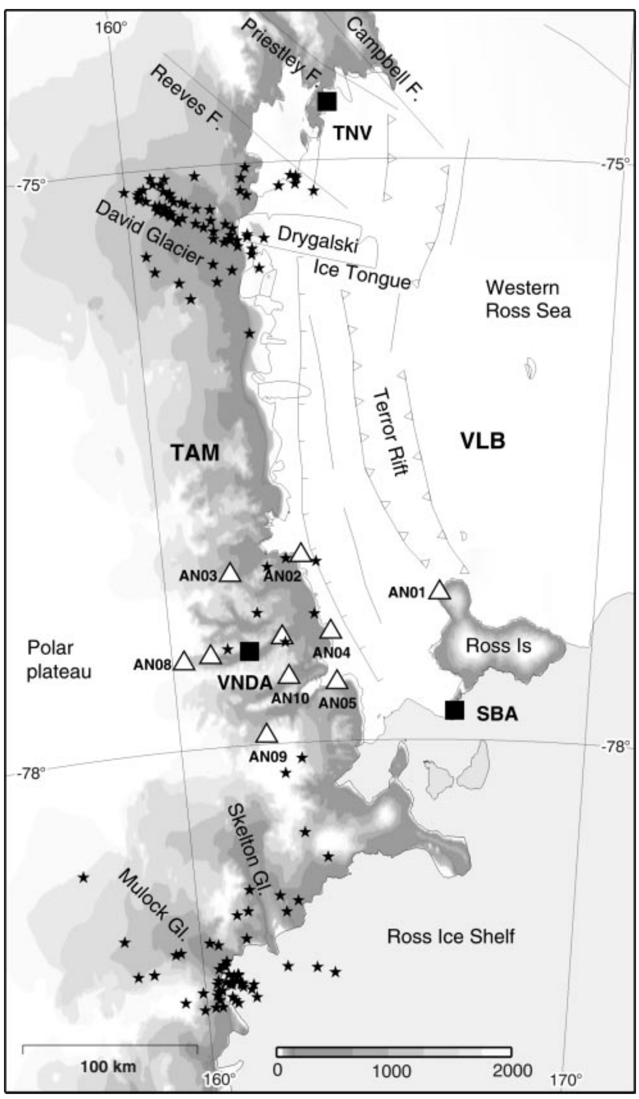

Fig 1 - The western Ross Sea region of Antarctica. Temporary seismic stations are shown as open triangles. Permanent seismograph stations at Vanda (VNDA), Scott Base (SBA) and Terra Nova Bay (TNV) are shown as solid squares. Seismic events located in this study are shown as stars. Over 60% of the events were located in the vicinity of the David glacier, and most of the remainder to the south, near Mulock glacier. Grey shading indicates elevation (in metres); VLB = Victoria Land Basin; TAM = Transantarctic Mountains.

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(17-22 km) crust (Trehu et al., 1989; O'Connell & Stepp, 1993; Cooper et al.,1987), relatively high heat flow, and the chemistry of alkaline lava. Within the VLB lies the Terror Rift, a zone involving recent (Neogene and younger) faulting as well as magmatic activity. Present-day alkaline volcanism is represented by Mt Erebus, located on Ross Island (Fig. 1) in the Ross Archipelago (LeMasurier, 1990), at the southern end of the Terror Rift. Onshore, Cenozoic tectonic faulting has been inferred in the Terra Nova Bay region (Salvini et al.,1997; Mazzarini et al.,1997; Salvini and Storti,1999), as well as in southern Victoria Land (Wilson,1999; Jones,1996). Although seismicity levels in Antarctica are regarded as low, active seismicity has been recorded in the western Ross Sea region and in the Transantarctic Mountains (TAM, Fig. 1) by several temporary array studies (*e.g.* Cimini et al., 1995; Privitera et al., 1992) and by permanent stations (Reading,1999). In this study we further examine the seismic activity in the TAM region, using digital data recorded by a temporary broadband seismometer array.

SEISMIC ARRAY DEPLOYMENT AND DATA ANALYSIS

During the 1999-2000 austral summer 10 broadband seismometer stations were deployed in the central Transantarctic Mountains (Fig.1), primarily to investigate the seismicity in the region, but also to provide data for crustal and upper mantle structure studies. The seismometer array, with an average station spacing of 20-25 km, straddled the Transantarctic Mountains opposite Ross island, with array stations near the edge of the plateau (*e.g.* station AN08, Fig. 1) as well as over the Victoria Land Basin (station AN01, Fig. 1). Each array station incorporated either a Guralp 40T or 3ESP seismometer, with an accompanying digital Orion recorder encased in a heavily insulated site-box, and with site power provided by solar panels.

Seismic data were also recorded by the permanent seismograph station VNDA (Fig. 1), which is located down a borehole in the Bull Pass region, the station SBA, located near Scott Base, and by the Italian broadband seismograph station TNV (Fig. 1), which is located at Terra Nova Bay (Morelli et al., 1994).

Event trigger algorithms were run through all of the broadband array data following data retrieval, as well as on the data from VNDA station, to search for, and associate, seismic events. Event hypocentres were subsequently determined using the Hypoellipse algorithm of Lahr (1999) based on arrival times picked from data from VNDA, TNV and the broadband array. One hundred and sixty seismic events were located, of which about 50 had horizontal position error estimates less than 20 km. Arrival time data from the TNV station were especially useful for constraining event locations. The distance of the events from the broadband array and VNDA station precludes accurate information on event depth. All event magnitudes were less than M_L3.5.

OBSERVATIONS

REGIONAL EVENTS TO THE NORTH

Eighty-four of the recorded seismic events were located 200-300 km to the north of the array, slightly inland from the Ross Sea coastline, in the vicinity of David Glacier and the Drygalski Ice Tongue (Fig. 1), distributed with a distinct ESE-WNW trend (Fig. 1). The recorded signal for these events involved distinct P and S phase arrivals, signal frequencies ranging between 1 and 9 Hz, and a signal duration averaging 60-100s, consistent with the seismic signals expected from normal tectonic events at that distance range. We discuss various alternative source mechanisms for these events further below.

These northern events are close in position to a postulated fault, variously termed the David fault (Salvini & Storti,1999) and the David Glacier lineament (Mazzarini et al.,1997). The right-lateral Reeves and Priestley faults (Salvini & Storti, 1999) lie just to the north of the events (Fig. 1). Salvini & Storti (1999) highlight the importance of these right-lateral strike-slip faults for the Cenozoic tectonic framework in the region.

Seismic events have also been observed in this general region, near David Glacier and the Drygalski Ice Tongue, in previous seismological studies by Privitera et al (1992) and Cimini et al (1995). Adams (1969, 1972) also located earthquake epicenters in the region, using readings from the early permanent stations at Scott Base (SBA) and Vanda (VNDA).

REGIONAL EVENTS TO THE SOUTH

Sixty-three seismic events were located southwest of Ross Island, to the south of the array, near the TAM Front at the eastern end of Mulock glacier (Fig. 1). Figure 2 shows the waveforms from one of these events, as recorded by VNDA station and 6 of the array stations. These southern events are tightly clustered near the transition between the Mulock glacier and the Ross Ice Shelf. The events lie close to a set of west-northwest-trending lineaments noted by Wilson (1999), which lie along and to the north of Mulock Glacier, at the southern boundary of an inferred rift accommodation zone (Wilson, 1999).

The recorded signals from the seismic events, with distinct P and S phase arrivals (Fig. 2) and signal frequencies ranging between 1 and 9 Hz, are consistent with seismic signal expected from earthquakes at distances around 100 km.

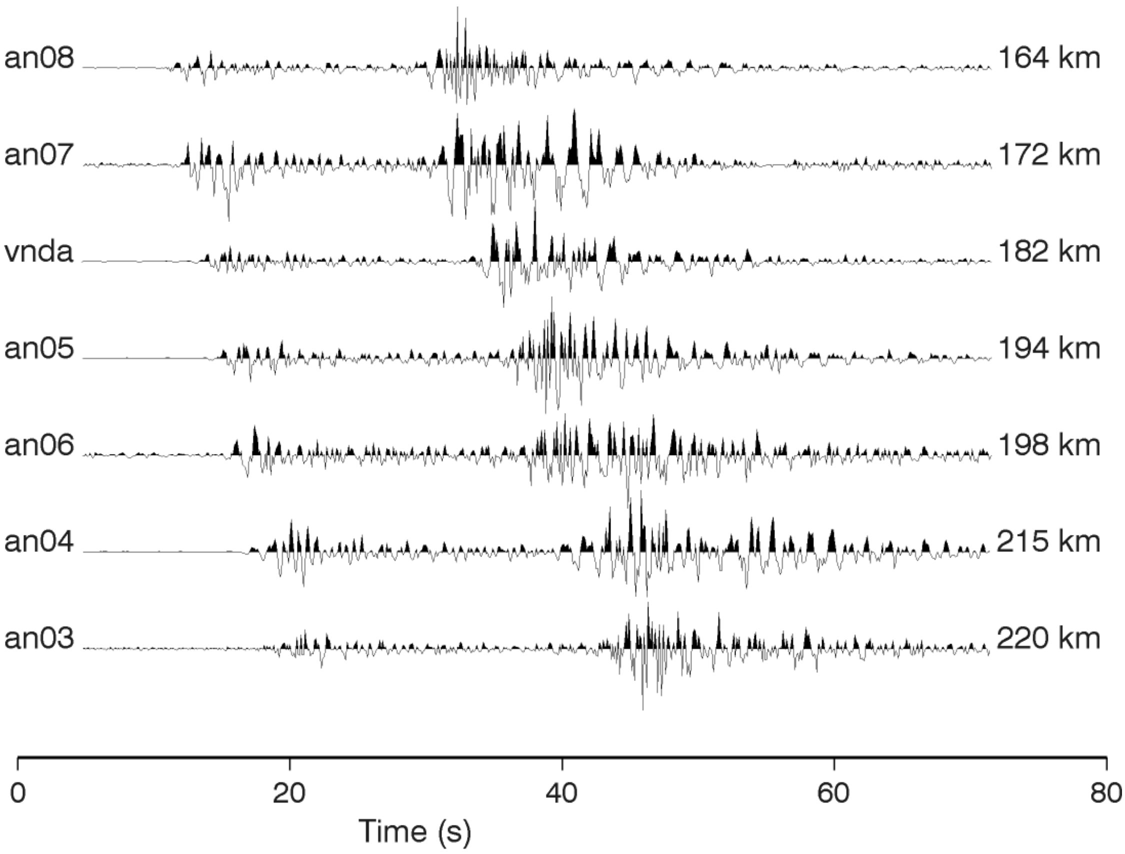

Fig 2 - Waveform data from a typical regional event recorded by the array stations. This particular event was located to the south of our array, on the coast near the Mulock glacier, south of the Skelton glacier. Event-station distances are annotated on the right.

LOCAL EVENTS

Eleven small seismic events were located inside the spatial extent of the broadband array (Fig. 1). The recorded waveforms from most of these local events have a high spectral content, and are quite impulsive, often with only a single clear phase arrival. These events were usually only recorded by a few stations, probably because the high frequency signal was quickly attenuated. The signal characteristics are similar to those associated with icequakes (*e.g.* Neave & Savage, 1970; Privitera et al., 1992) rather than earthquakes.

DISCUSSION

ALTERNATIVE SOURCE MECHANISMS

Seismic events recorded in Antarctica have variously been attributed to icequakes resulting from brittle ice deformation on glaciers (*e.g.* Privitera et al., 1992), calving of icebergs (Hatherton & Evison, 1962), basal-shearing beneath glaciers (Anandakrishnan & Alley, 1997), and tectonic earthquakes occurring along faults (*e.g.* Kaminuma & Akamatsu,1992). The signal characteristics from each of these possible source mechanisms will differ, as may the expected level of seismic energy release.

Seismic signals are known to be generated by iceberg calving near the termini of large glaciers (*e.g.* Qamar, 1988). Such seismic signals have emergent onsets lacking a distinct arrival, a weakly developed P phase, an obscured S arrival followed by a non-dispersive wave train, and a monochromatic low frequency (1-2 Hz) signature associated with both P and S phases (Wolf and Davies, 1986; Qamar,1988; Gaull et al,1992). In one study of ice calving Qamar (1988) noted that virtually every calving event produced this low frequency seismic signal, with a signal duration corresponding to the duration of the calving, between 13 and 150s. They found that the recorded signals from the calving events were distinctly different from recordings of local earthquakes. Seismic events observed in our study do not have the signal characteristics noted by ice-calving studies. In addition, most of our seismic events are located well inland, away from glacier termini, so we discount this source mechanism for our events.

Local “icequakes” near the surface of glaciers, assumed to represent brittle ice deformation (Sprenke et al., 1997), have a seismic signature distinctly different from ice-calving events. In a detailed study of icequakes, Neave and Savage (1970) determined epicentres for more than 1000 events recorded on the Athabasca glacier, in Canada. They found that the icequakes generally occurred in the marginal crevasse zones of the glacier, and that, from first motion analysis, there was little doubt that the icequakes were associated with extensional faulting (crevassing) rather than stick-slip faulting. In our study region, the David glacier involves a large elevation drop and has visible surface crevasses, indicating that brittle ice deformation can reasonably be expected, possibly generating icequakes. Icequakes have already been observed in the David Glacier area by Privitera et al (1992), who noted that events recorded by seismometer stations on the glacier itself were not recorded by stations only 2 km away; and that the seismic waves rapidly lost high frequency content. Characteristics of the recorded seismic signals from these icequake events included weak onset, no evidence of S phase, a sharp decrease in amplitude with time, a high spectral content (> 10 Hz), and a short signal duration (Privitera et al., 1992). We note that these signal characteristics are quite different from the regional waveforms observed in our study.

Icequakes could also potentially be generated at steep ice falls, from the vertical movement of large volumes of ice (*e.g*. Weaver & Malone, 1979); ice volumes on the order of 10^4 m^3, instantaneously moving 0.01m vertically, with a 1% energy-conversion efficiency, could potentially produce $O(10^5$ J) of seismic energy.

Stick-slip motion at the base of a glacier is another recognised source for seismic signals (*e.g*. Weaver & Malone, 1979; Anandakrishnan & Langston, 1994; Anandakrishnan & Alley, 1997). Slip event occurrence depends on the nature of the shear stress in the basal region, which will vary with ice thickness, the nature of the sub-glacial till, and the glacier-surface slope (*e.g.* Paterson, 1981). In one study Anandakrishnan and Langston (1994) located low-angle thrust-fault slip events at the bed of West Antarctic ice stream-B, with fault plane areas $O(10^2$ $m^2)$. In the present study we note that if our seismic events represent basal-slip events then, depending on the energy conversion efficiency, the fault plane area would need to be $O(10^3$ $m^2)$ to release the $O(10^7$-10^9 J) necessary for the observed events (following Richter, 1958).

Alternatively, the seismic events observed in the David glacier region could well represent earthquakes generated by deep-seated faulting beneath the glacier. The presence of a large regional lineament beneath the David Glacier, the "David Glacier lineament", was inferred by Mazzarini et al (1997), who suggested that it represented a major transfer structure across the TAM. Their interpretation was primarily based on differences in structural trends, glacial patterns, uplift magnitudes and geophysical characteristics north and south of the lineament.

The presence of a major crustal structural discontinuity beneath David Glacier was similarly suggested by Redfield and Behrendt (1992), on the basis of gravity observations across the David Glacier, although their data are not conclusive. Bozzo et al. (1997) noted a prominent WNW-ESE aeromagnetic anomaly in the same area, which has the same trend as the seismic event distribution. The seismic waveforms recorded by our stations from the seismic events are quite consistent with the seismic signal expected from a regional earthquake at O(200 km) distance, with seismic frequencies ranging between 1 and 9 Hz.

If the seismic events do represent earthquakes, then we can expect that their distribution will be affected by post-glacial rebound, as well as by regional tectonic forces (*e.g*. Wu et al., 1999; Muir-Wood, 2000). Modelling studies have shown that, although rebound-induced stress changes may be too small to cause fracture, they may be enough to trigger earthquakes on pre-existing faults, depending on the regional stress-field orientation (Wu et al., 1999). However, more information is needed on the earthquake source mechanisms, uplift rates, time and spatial variation of the ice model, and the regional stress field before we fully understand the effect of isostatic rebound on seismicity patterns in the Western Ross Sea region.

SOURCE MECHANISM DISCRIMINATION

The source mechanism for a seismic event may be distinguished if additional information such as the source depth or the focal sphere mechanism can be determined. For example, basal events can be distinguished from surface crevassing events if the recording stations are close to or above the events (*e.g*. Anandakrishnan & Alley, 1997), on the basis of the event depth and mechanism.

At regional distances Rayleigh wave (Rg) excitation known to be very dependent on source depth; Rg is usually absent at regional distances if the source depth is greater than a few kilometres (Xie and Lay, 1994). We note that the ratio of Rg to shear wave energy recorded at station TNV is relatively low for many of the events observed near David Glacier, indicating a reasonable source depth. However, for an unambiguous interpretation we still require some 'calibration' earthquakes, with good estimates of hypocentral depth. Secondary phases such as pP and PmP, when correctly identified, can also be used for improving estimates of hypocentral depth. Polarisation analysis of records from some of the events near David Glacier show secondary phases, 1 to 2s after the P-arrival, with steep incident angles. If these phases are interpreted as surface reflection pP arrivals then reasonable source depths, O(10 km), are implied for those events.

Source mechanism discrimination also requires quality focal sphere coverage, in addition to accurate hypocentres. Basal-thrust-slip events and (surface) extensional ice deformation events could potentially be distinguished on the basis of fault-plane solutions (e.g. Sprenke et al.,1997). We would hope to collect such quality focal sphere information in future targetted deployments, to assist source discrimination.

CONCLUSIONS

Local and regional seismic events have been recorded in the Transantarctic Mountains using a 10-station broadband seismometer array. Events located in the vicinity of David Glacier, inland from the Drygalski Ice Tongue, are aligned ESE-WNW, close to a proposed David Glacier lineament, and sub-parallel to the Reeves and Priestley faults. The available data do not however allow unique discrimination between possible deep-seated faulting along the lineament and other source mechanisms such as brittle ice deformation on the surface of the David glacier. However, we discount iceberg calving at the David glacier termini as a source mechanism, on the basis of the signal characteristics and the event distribution.

Events to the south of our array were located very close to the transition between the Mulock glacier and the Ross Ice Shelf. These events may represent either basal stick-slip movement beneath the Mulock glacer, brittle ice deformation at the glacier transition, or tectonic earthquakes associated with west-northwest trending lineaments noted by Wilson (1999) in that area.

Although we cannot uniquely constrain the source mechanisms of the seismic events, we present these data in order to assist further investigations. Additional analyses are planned to examine whether the available data provide any further information on event depth.

Clearly, one of the ongoing priorities for seismological research in Antarctica is source mechanism discrimination for observed seismic events. We expect new targeted experiments will help to resolve some of the issues, especially as new permanent seismic observatories and temporary seismic arrays are deployed, and as magnitude detection levels are lowered.

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REFERENCES

Adams R.D., 1969. Small earthquakes in Victoria Land, Antarctica, *Nature*, **224**, 255-256.

Adams R.D., 1972. Local earthquakes in Victoria Land. *Antarctic geology and geophysics*, **NUMBER?**, 495-499.

Anandakrishnan S. & Alley R.B.,1997. Tidal forcing of basal seismicity of ice stream C, West Antarctica, observed far inland. *Journal of Geophysical Research*, **102**, 15183-15196.

Anandakrishnan S. & Langston C.A.,1994. Seismology on ice: Source locations of microearthquakes under ice streams B and C, West Antarctica. *Seismological Research Letters*, **65**, 62.

Bozzo E., Ferraccioli F., Gambetta M., Caneva G., Damaske D., Chiappini M. & A.Meloni, 1997. Aeromagnetic regional setting and some crustal features of Central-Southern Victoria Land from the GITERA surveys. In: C.A.Ricci (ed.), *The Antarctic Region: Geological Evolution and Processes*, Terra Antartica Publication, Siena, 591-596.

Cimini G.B., Amato A., Cerrone M., Chiappini M., Di Bona M., and S.Pondrelli, 1995. Passive seismological studies in the Terra Nova Bay area (Antarctica): First results from the 1993-94 expedition, *Terra Antartica*, **2**, 81-94.

Cooper A.K., Davey F.J. & Behrendt J.C., 1987. Seismic stratigraphy and structure of the Victoria Land Basin, Western Ross Sea, Antarctica. In: Cooper A.K. & Davey F.J. (eds.), *The Antarctic Continental Margin: Geology and Geophysics of the Western Ross Sea*, Earth Science Series **5B**, Houston, Texas, Circum-Pacific Council for Energy and Mineral Resources, 27-66.

Gaull B.A., Adamson D.A. & Pickard J., 1992. Seismicity associated with icebergs calving from glaciers near Mawson, East Antarctica, *Australian Journal of Earth Sciences*, **39**, 473-480.

Hatherton T. & Evison F.F., 1962. A special mechanism for some antarctic earthquakes, *N.Z. Journal Geol. Geophys.*, **5**, 864-873.

Jones S., 1996. Late Quaternary faulting and neotectonics, South Victoria Land, Antarctica. *J. Geol. Soc., London*, **153**, 645-652.

Kaminuma K. & Akamatsu J.,1992. Intermittent micro-seismic activity in the vicinity of Syowa station, East Antarctica. In: Y. Yoshida et al.(eds.), *Recent Progress in Antarctic Earth Science*, Terra Scientific Publishing Company, Tokyo, 493-497.

Lahr J.C., 1999. HYPOELLIPSE: A computer program for determining local earthquake hypocentral parameters, magnitude, and first motion pattern, (Y2K compliant version) *U.S.Geological Survey Open-File Report 99-23*, 112p.

LeMasurier W., 1990. Late Cenozoic volcanism on the antarctic plate – an overview. In: LeMasurier W., Thompson J.W., Baker P.E., Kyle P.R., Rowley P.D., Smellie J.L. & Verwoerd W. J. (eds.), *Volcanoes of the Antarctic plate and Southern Oceans AGU Antarctic Research Series*, **48**, 1-17.

Mazzarini F., Della Vedova B. & Salvini F., 1997. Crustal segmentation of the Transantarctic Mountains Rift Shoulder along the David Glacier Lineament, Victoria Land (Antarctica). In: Ricci,C.A. (ed.), *The Antarctic Region: Geological Evolution and Processes*. Terra Antartica Publication, Siena, 565-569.

Morelli A., Romeo G., Meloni A. & Palangio P., 1994. Contribution of the permanent seismograph station at Terra Nova Bay to global seismological research. *Terra Antartica*, **1,** 184.

Muir-Wood R., 2000. Deglaciation seismotectonics: a principal influence on intraplate seismogenesis at high latitudes, *Quaternary Science Reviews*, **19**, 1399-1411.

Neave,K.G., and J.C.Savage, 1970. Icequakes on the Athabasca glacier, *Journal of Geophysical Research*, **75**, 1351-1362.

O'Connell D.R.H & Stepp T.M., 1993. Structure and Evolution of the crust at the Transantarctic Mountains - Ross Sea crustal transition: Results from the Tourmaline plateau seismic array of the GANOVEX V Ship-to-shore seismic refraction experiment. *Geologisches Jahrbuch*, **47**, 229-276.

Okal E.A., 1981. Intraplate seismicity of Antarctica and tectonic implications. *Earth and Planetary Science Letters*, **52**, 397-409.

Paterson W.S.B., 1981. *The Physics of Glaciers*, Elmsford, New York. Pergamon Press.

Privitera E., Villari L. & Gambino S., 1992. An approach to the seismicity of Mt. Melbourne volcano (Northern Victoria Land - Antarctica). In: Y.Yoshida et al. (eds.), *Recent Progress in Antarctic Earth Science*, Terra Scientific Publishing Company, Tokyo, 499-505.

Qamar A., 1988. Calving icebergs: a source of low-frequency seismic signal from Columbia Glacier, Alasak. *Journal of Geophysical Research*, **93**, 6615-6623.

Reading A.M., 1999. Antarctic seismicity and neotectonics. In: Gamble J.A., Skinner D.N.B., Henrys S. & Lynch R. (eds.) *"Antarctica at the close of a Millenium" Proceedings Volume 8th International Symposium on Antarctic Earth Sciences*, Royal Society of NewZealand Bulletin, **35** (in press).

Redfield T.F. & Behrendt J.C., 1992. Gravity modelling across the Transantarctic Mountains, Northern Victoria Land. In: Y. Yoshida et al. (eds.), *Recent Progress in Antarctic Earth Science*, Terra Scientific Publishing Company, Tokyo, 535-544.

Richter C.F., 1958. *Elementary seismology*, San Francisco, W.H.Freeman and Co.

Salvini F. & Storti F., 1999. Cenozoic tectonic lineaments of the Terra Nova Bay region, Ross Embayment, Antarctica. *Global and Planetary Change*, **23**, 129-144.

Salvini F., Brancolini G., Busetti M., Storti F., Mazzarini F. & Coren F., 1997. Cenozoic right-lateral strike-slip tectonics in the Ross Sea region, Antarctica. In: C.A.Ricci (ed.), *The Antarctic Region: Geological Evolution and Processes*, Terra Antarctica Publication, Siena, 585-590.

Sprenke K.F., Adema G.W. & Miller M.M., 1997. Brittle ice deformation on the Juneau Icefield, an analog for icequake activity on other planetary bodies. *Eos, Trans.*, American Geophysical Union, **78** (46, Suppl.), 409.

Trehu A.M., Holt T., Behrendt J.C. & Fritsch J.C., 1989. Crustal structure in the Ross Sea , Antarctica: preliminary results from GANOVEX V (abstract), *EOS Trans.*, American Geophysical Union, **70**, 1344.

Weaver C.S. & Malone S.D., 1979. Seismic evidence for discrete glacier motion at the rock-ice interface, *Journal of Glaciology*, **23**, 171-184.

Wilson T.J., & Finn C.A. (eds.), 1996. *Geodynamic Evolution of the Transantarctic Mountains and the West Antarctic Rift System. Proceedings of a Workshop*. BPRC Report Byrd Polar Research Center, Columbus, OH, 57pp.

Wilson T.J., 1999. Cenozoic structural segmentation of the Transantarctic Mountains rift flank in southern Victoria Land. *Global and Planetary Change*, **23**, 105-127.

Wolf L.W. & DaviesJ.N., 1986. Glacier generated earthquakes from Prince William Sound, Alaska. *Bull. Seism. Soc.* ***Am.***, 76, 367-379.

Wu P., Johnston P. & Lambeck K., 1999. Postglacial rebound and fault instability in Fennoscandia. *Geophys. J. Int.*, **139**, 657-670

Xie X.B. & Lay T., 1994. The excitation of Lg waves by explosions: a finite difference investigation, *Bull. Seism. Soc. Am.*, **84**, 324-342.