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THE BELLINGSHAUSEN SEA EARTHQUAKE OF FEBRUARY 5, 1977: EVIDENCE FOR RIDGE-GENERATED COMPRESSION IN THE ANTARCTIC PLATE

EMILE A. OKAL

Department of Geology and Geophysics, Box 2161, Yale University, New Haven, CT 06520 (U.S.A.)

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Body- and surface-wave data from a magnitude $M_s = 6.4$ event in the Bellingshausen Sea are used to infer a thrust fault mechanism, with compressional stress directed parallel to the vector of relative motion between Antarctica and the Pacific plate at the nearby ridge. The solution would not, on the other hand, be compatible with stress fields involving absolute plate motion, such as asthenospheric drag.

1. Introduction

Among the Earth's plates, Antarctica is unique in that it is almost entirely surrounded by ocean ridges. Only along the southern tip of South America and the South Scotia Sea are different types of boundaries (primarily transform faults) found. Antarctica is thus a relatively simple proving ground for dynamic models of plate motion, especially in the wake of recent observations [1,2] emphasizing the importance of ridge push in the balance of forces driving the plates. Antarctica is also unique as a continent in its low seismicity. Over the past 15 years, only a dozen or so events with $m_b \geqslant 4.5$ were located inside the plate, and this led Sykes [3] to hypothetize that a quasi-continuous ring of spreading centers surrounding the continent, together with a low absolute velocity with respect to the mantle, may impede the transmission of tectonic stress into the plate, either from below or from adjacent plates. The interior of the plate may then be viewed as being in a passive state, while the ridges simply move away from it. Seismic events in the Antarctic plate therefore deserve special attention.

A 1971, $m_b = 6.2$ shock studied by Forsyth [4] occurred in the plate's extreme northern corner, between the two fast-spreading East Pacific and Chile ridges. The thrust mechanism obtained by Forsyth indicated horizontal compressional stress generated by ridge push. This isolated event may, however, have represented a very localized state of stress, due to the immediate vicinity of the two ridges, and not be representative of the stress condition in the interior of the plate.

The recent occurrence of a second large event inside the Antarctica plate, this time far away (2900 km) from the nearest ridge system, may drastically change our views on Antarctic seismicity. The study of this $m_b = 6.2$ shock in the Bellingshausen Sea is the subject of the present paper. This earthquake is of particular interest, since its mechanism reflects the state of stress in the lithospheric plate away from its boundaries.

2. Seismic investigation

The Bellingshausen Sea earthquake occurred at 0.3 : 29 GMT on February 5, 1977, and was located by the International Seismological Centre (ISC) at 66.5°S, 82.4°W, in the gently sloping continental rise lying between the Bellingshausen Abyssal Plain and the Continental Shelf (see Fig. 1). This area is one of very complex past tectonics, involving subducted

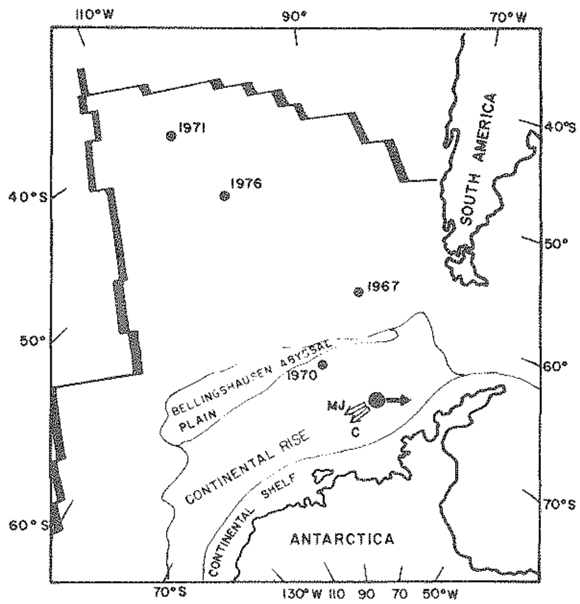

Fig. 1. Map of the southeastern Pacific, showing plate boundaries, and major bathymetric features. Intra-plate epicenters in the area are shown as full circles. The 1971 event was studied by Forsyth [4]. The larger symbol shows the Bellingshausen Sea 1977 event. Also shown are the directions of the local velocity vectors, both for relative motion of the Pacific and Antarctic plates (full arrow) and for absolute motion (open arrows). In this latter case, *MJ* and *C* refer to Minster and Jordan's [14] and Chase's [15] models, respectively.

spreading centers and episodes of ridge jumping [5,6]. The local bathymetry [7] indicates an average depth to the seafloor of 2300 fathoms (4200 m).

No aftershocks to this event were reported, but a steady, albeit low seismicity in the Humboldt and Bellingshausen plains is documented by two events in 1967 ($m_b = 5.3$) and 1970 ($m_b = 4.9$) (see Fig. 1). Little information is available on the depth of the event, given by NEIS as 33 km, and by ISC as 31 ± 19 km from P-wave observations, and 40 ± 1 km from pP data. This last estimate is particularly suspect, since the ISC itself lists practically no pP observations. High-gain short-period vertical records at Mundaring, Caracas and Otepa clearly show a depth phase (11.5 ± 1) seconds after P. This would correspond to a depth of (36 ± 4) km, if we interpret it as pP, and according to the J-B Tables. On the other hand, if the reflection took place at the ocean's surface, the phase is then pwP [8], and the real focal depth would be much shallower (15 km). However, in this case one could expect to observe pP before pwP; also the successful input of elastic energy into the water column necessary to excite pwP usually results in numerous water multiples, such as pwwP and pwwwP [8,9], with an expected time separation of twice the one-way vertical travel-time t up the water column (in the present case $2t = 5.5$ seconds); neither of these phases are present in the records. We conclude that the correct interpretation of the observed phase is pP, and in analyzing first-motion data, we use a focal depth of 35 km with a P-wave velocity of 8.1. km/s. We will see that surface-wave data confirm this depth. Since most WWSSN stations with high magnifications were either at distances larger than 90°, or close to a nodal plane, it was not possible to use long-period body-wave shapes to further constrain the focal depth.

Body-wave first motion data is shown in Fig. 2. All P-wave first motions read for the present study were compressional, as were all but one of the first motions reported to the ISC, relative to direct P, and with residuals smaller than 2.5 seconds. S-wave directions of motion at a number of stations are also shown in Fig. 2, indicating high values of the amplitude ratio SV/SH. This set of first-motion data clearly demonstrates the thrust nature of the faulting, but fails to further constrain the mechanism.

For this purpose, we use a set of 22 Love and 20 Rayleigh-wave records from mostly WWSSN stations, well distributed in azimuth (see Table 1). We follow the equalization technique of Kanamori [10], after low-pass filtering the data with a cut-off frequency of 0.022 Hz, and a linear taper down to 0.016 Hz. As a result, the dependence of azimuthal radiation patterns on the poorly known depth of the event, and on lateral structural heterogeneities along the path of the waves, is greatly reduced. The OTP records were also equalized to the WWSSN instrument response. The resulting radiation patterns are shown in Fig. 3. Synthetic seismograms built through normal mode summation for simple mechanisms were then linearly combined for a systematic computer search of the focal solution fitting the surface-wave data best, while remaining compatible with body-wave first motions. Three focal depths of 20, 35 and 50 km were tried, covering the range of uncertainty of the ISC solution.

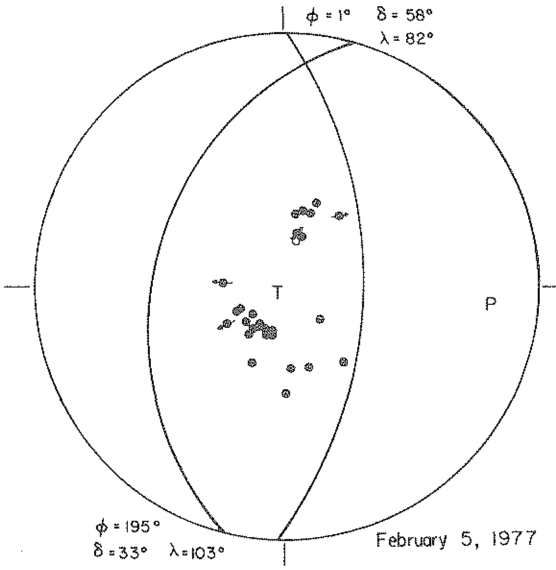

Fig. 2. Focal mechanism obtained from body- and surface-wave data. Open circles are dilatations, full circles compressional arrivals. The small arrows show the polarity of first motion of S waves. P = compression axis; T = tension axis. Both fault planes are constrained by surface waves.

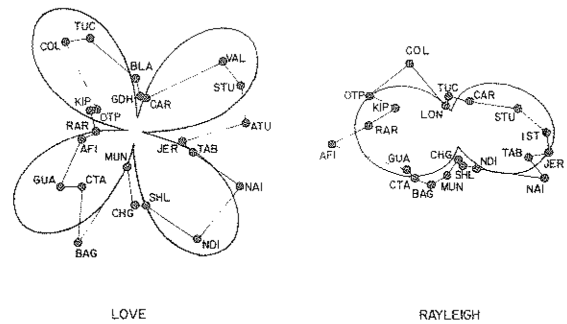

Fig. 3. Equalized surface-wave radiation patterns. Large dots are individual data points at the various stations. The continuous thick trace is the theoretical diagram representing the amplitude of synthetics built with the same filter characteristics, and a seismic moment of 4.4×10^{25} dynes-cm.

The overall Love-to Rayleigh ratio of about 0.8 was best fit by a depth of 35 km. This may appear surprising, since many intra-plate oceanic earthquakes have been found to be much shallower [4,8,9]. However, it was clear from our results that a depth of 20

TABLE 1

Seismic records used in the surface wave study

Code	Station	Epicentral distance (°)	Azimuth to station (°)	Waves used
BLA	Blacksburg, Virginia	103.55	1.7	G
GDH	Godhaab, Greenland	137.08	14.7	G
CAR	Caracas, Venezuela	77.73	15.7	G, R
VAL	Valentia, Ireland	130.17	50.6	G
STU	Stuttgart, Germany	134.13	67.2	G, R
ATU	Athens, Greece	130.59	87.0	G
IST	Istambul, Turkey	135.32	89.6	R
JER	Jerusalem, Israel	129.74	101.7	G, R
TAB	Tabriz, Iran	139.59	108.9	G, R
NAI	Nairobi, Kenya	100.09	117.8	G, R
NDI	New Delhi, India	140.18	151.7	G, R
SHL	Shillong, India	138.92	172.3	G, R
CHG	Chiengmai, Thailand	132.30	181.8	G, R
MUN	Mundaring, Australia	80.65	196.0	G, R
BAG	Baguio, Philippines	127.69	208.2	G, R
CTA	Charter Towers, Australia	86.23	224.9	G, R
GUA	Agana, Guam	118.52	234.3	G, R
AFI	Afiamalu, Western Samoa	77.05	264.9	G, R
RAR	Rarotonga, Cook Islands	65.64	272.6	G, R
KIP	Kipapa, Oahu, Hawaii	103.94	291.5	G, R
OTP	Otepa, Hao, French Polynesia	61.07	292.1	G, R
COL	College, Alaska	139.33	323.8	G, R
LON	Longmire, Washington	117.01	330.7	R
TUC	Tucson, Arizona	101.00	335.8	G, R

km, for example, would require much larger Rayleigh waves, and lead to poorer simultaneous fits of the two radiation patterns. The influence of the poorly known oceanic structure, and of such factors as conversion from oceanic to continental structure along the path, on the Love-to-Rayleigh ratio [8] is minimized by our filtering of the data, since most of the energy in the resulting signals is around 70 seconds (0.014 Hz). With a depth of 35 km, the solution converged in three iterations, to a precision of one degree. The final solution (strike N1°E; dip 58°; rake 82°) represents a nearly pure dip-slip along a reverse-faulting north-striking plane. This solution is almost identical to Forsyth's [4] mechanism for the 1971 event further north. It is also very similar to Mendiguren's [8] solution for a strong shock in the middle of the Nazca plate. The accuracy of the solution is about 5° in each of the angles and should not affect the conclusions of the paper. The seismic moment was found to be $(4.4 \pm 1.5) \times 10^{25}$ dynes-cm, and a reevalution of the surface-wave magnitude yielded $M_s = 6.4 \pm 0.3$ at 18 seconds. In the absence of any data on aftershocks, or surface of faulting, it is impossible to compute an estimate of the stress drop involved, but the above figures suggest an apparent stress $\eta\bar{\sigma}$ of about 20 bars, comparable to worldwide averages, notably for interplate earthquakes. This suggests that the distinction in stresses between interplate and intra-plate events [11] may be futile.

3. Discussion

The relative importance of boundary forces, such as ridge push or slab pull, and surface area forces, such as lithospheric drag, on the state of stress within a plate has long been debated [1,2,12]. Ridge push should give rise to compressional stresses oriented parallel to the *relative* motion of the two plates at the ridge, whereas stresses due to the drag should be oriented parallel to the direction of *absolute* motion of the plate, in a frame where the mantle remains fixed. Unfortunately, most previous studies of intraplate seismicity have delt with the North American plate [13] or the Pacific plate [9] or the South American plate [2], where these two directions are similar or very close, due to a particular geometry of the plate motions. This makes any discrimination between the two interpretations of the seismicity impossible. The present event is, however, perfectly located for a test of these models, since the directions of relative and absolute motions are well separated at the epicenter: we computed the azimuths of local directions of motion both from Minster and Jordan's [14] and Chase's [15] models. In both models, the relative motion between the Pacific and Antarctic plates strikes N91°E; the absolute motion with respect to the mantle strikes N119°W in Minster and Jordan's model, N130°W in Chase's. As shown in Fig. 2, the compressional axis obtained from our mechanism strikes N96°E, only 5° away from the direction of relative motion, but 40° away from that of absolute motion (see (Fig. 1). It would be impossible to rotate our mechanism 40°, or to change its nature, without drastically altering the surface-wave radiation patterns. It could be argued, however, that the Antarctic plate is found in both models to have a very slow absolute motion, and that the direction of this motion may be poorly determined. This would in turn argue against interpreting seismicity in the Bellingshausen Sea as being due to asthenospheric drag, since in this case shear stresses at the base of the plate would also vanish proportionally to the absolute velocity. We must therefore conclude that the stress released during the 1977 Bellingshausen Sea earthquake was due to ridge push from the spreading center along the Pacific-Antarctic ridge, rather than to a compressive state resulting from the drag of the plate on the asthenosphere below. This result extends Forsyth's [4] conclusions to an area more representative of the interior of the plate. It also agrees with Mendiguren and Richter's [2] conlcusions, although the geometry of the Antarctic plate would require a set of boundary conditions different from that used in their model for South America.

In the absence of any detailed bathymetric chart, it is difficult to associate the Bellingshausen Sea earthquake with any particular feature suggesting a pre-existing zone of weakness. The epicenter is located in an area of numerous fracture zones, belonging to past plate tectonics patterns, the lithosphere in the area having probably been generated about 55 m.y. ago, at a ridge which separated East and West Antarctica until the Eocene (this ridge has since been consumed in the subduction zone which probably existed along the Antarctic Peninsula [5]).

It would therefore be incorrect to try to interpret the weakness around the epicenter in the pattern of presently active fracture zones.

4. Conclusion

The Bellingshausen Sea earthquake of February 5, 1977, clearly demonstrates that the intra-plate seismicity in the Antarctic plate is not ananomalously low, as previously believed. The focal mechanism obtained from body- and surface-wave data is of the thrust fault type, with the compressional axis in the direction of ridge push at the nearby Pacific-Antarctic boundary. This mechanism would not be compatible with stresses generated by asthenospheric drag at the base of the plate. This indicates that tectonic stresses are indeed transmitted into Antarctica through the surrounding ridges, and that stresses due to ridge push are responsible for its intra-plate seismicity. Significantly, the only portion of the Antarctic plate which showed substantial seismicity ($m_b \geqslant 5.0$) over the past 15 years is that closest to the Pacific ridge, which has the fastest spreading rate of all ridges surrounding Antarctica, and therefore creates the strongest stresses in the plate.

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