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On moment-length scaling of large strike slip earthquakes and the strength of faults

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[1] Several large strike slip earthquakes have occurred in various tectonic settings in the past 5 years, adding well documented data to the global collection of moment and length estimates for such earthquakes. Based on this augmented dataset, we reexamine the controversial issue of scaling of seismic moment with length of rupture. We find that the global dataset of large strike-slip earthquakes follows a bi-modal distribution. Most oceanic and/or intraplate strike-slip earthquakes have stress drops that are ~5 times larger than interplate continental ones. When distinguishing these two classes, the scaling is compatible with that predicted by simple dislocation theory. *INDEX TERMS:* 7209 Seismology: Earthquake dynamics and mechanics; 7215 Seismology: Earthquake parameters; 7230 Seismology: Seismicity and seismotectonics

1. Introduction

[2] There has been a long lasting controversy in the literature as to whether earthquake moment (Mo) scales with L^2 or L for large earthquakes, where L is the length of the fault. In simple terms, the issue hinges on whether the average slip d during an earthquake grows with the length L or the width W of the fault. The issue of scaling is particularly important for seismic hazard estimation based on lengths of fault segments, since significantly different estimates of maximum possible earthquake size can be obtained for a given region, depending on the scaling law.

[3] We start from the definition of seismic moment: $M_0 = \mu dLW$, where μ is shear modulus. For small earthquakes, for which $W < W_o$ (the maximum width allowed by the thickness h of the brittle zone), it is generally assumed that the rupture grows in both L and W, so that $L = fW$, where f is a geometrical factor. In this case, either model predicts that M_o should scale with L^3, or with $S^{3/2}$, where S is the fault area, in agreement with observations [e.g. *Kanamori and Anderson*, 1975]. When $W = W_o$, a change of scaling occurs, as the rupture can, from then on, grow only in one dimension. The issue of scaling for large earthquakes is best addressed from observations of strike-slip earthquakes on quasi-vertical transcurrent faults, since in that case $W_o \sim h$ and $h \sim 15-25$ km. This removes the additional degree of freedom arising from the large variability in the dip, and therefore W, of large thrust and normal faulting events.

[4] Dislocation theory predicts that stress drop $\Delta\sigma$ is proportional to d/W, hence slip scales with W for constant $\Delta\sigma$. This implies scaling of M_0 with L^n, where $n = 3$ for small earthquakes and $n = 1$ for large earthquakes with $W = W_0$. *Scholz* [1982] proposed an alternative model, in which the slip scales with L. This model was motivated by inspection of slip versus length data that were available at that time. It implied that $n = 2$ for large earthquakes. On the other hand, *Romanowicz* [1992] compiled the existing dataset for large strike-slip earthquakes on quasi-vertical transcurrent faults. She concluded that moment scales with $n = 3$ for moments smaller than $\sim 0.6-0.8 \times 10^{20}$ Nm, as known previously, while for larger moments, the data favored a scaling with $n = 1$, compatible with dislocation theory. *Romanowicz and Rundle* [1993] then showed, based on scale invariance arguments [e.g. *Rundle*, 1989], that the $n = 1$ and $n = 2$ scalings could also be differentiated on the basis of frequency-moment statistics, favoring of the "W-model".

[5] Since then, the controversy has continued, using theoretical [e.g. *Sornette and Sornette*, 1994; *Romanowicz*, 1994; *Bodin and Bilham*, 1994] as well as observational arguments, the latter mostly involving compilations of M_0 versus L [e.g. *Scholz*, 1994; *Pegler and Das*, 1996] but also waveform modeling of source finiteness [*Mai and Beroza*, 2000]. On the other hand, new compilations of slip versus length data indicate that the increase of slip with L tapers off at large L [e.g. *Bodin and Brune*, 1996]. This view has recently received further support from numerical modelling [*Shaw and Scholz*, 2001]. Recently, *Miller* [2002] proposed a scaling which depends on fault zone pore pressure.

[6] *Romanowicz* [1992] classified several recent large strike-slip earthquakes that occurred in oceanic tectonic settings (Macquarie, 1989; Alaska '87–'88) as unusual, in that their scaling did not agree with the $n = 1$ model. Detailed studies have been conducted since, yielding more precise estimates of fault length for these earthquakes. Furthermore, in recent years, several very large strike-slip earthquakes have occurred in different tectonic environments, and some of them have been extensively studied. This has motivated us to re-examine the issue of scaling of M_0 with L.

2. Dataset and Observed Trends

[7] In general, M_0 estimates are much more accurate than those of L. Since 1977, the Harvard CMT catalog [*Dziewonski et al.*, 1981] provides robust M_0 estimates from seismic waveforms for earthquakes larger than $M \sim 5.5$. For events of magnitude larger than $M7$, the catalog compiled by *Pacheco and Sykes* [1992] takes us back to the beginning of the 20th century. On the other hand, L is mostly estimated from the distribution of aftershocks, except for continental earthquakes for which additional constraints are obtained from surface rupture observations. In general, the latter method leads to an underestimation of rupture length, and the former, to an overestimation.

[8] We consider the catalog of *Pegler and Das* [1996] (PD96 in what follows), who have combined M_0 estimates from the Harvard CMT catalog, with L for large earthquakes from 1977 to 1992 based on relocated 30-day aftershock zones. We add to this dataset the standard collection of reliable M_0/L data for large strike-slip earthquakes since 1900 [e.g. *Romanowicz*, 1992], data for great central Asian events since the 1920's [*Molnar and Denq*, 1984], as well as data for recent large strike-slip events (e.g. Balleny Islands '98; Izmit, Turkey '99 and Hector Mines, CA, '99) that have been studied using a combination of modern techniques (i.e. field observations, waveform modelling, aftershock relocation).

[9] We also consider 15 other strike-slip events of moment $M_o > 0.05 \times 10^{20}$ Nm that occurred in the period 1993–2001.

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Table 1. Dataset Used in This Study

Name	Date	Length	M_0	Type	Ref.	Name	Date	L	M_0	Type	Ref.
1 S. Francisco	04/18/1906	450	4.0	A	2	38 New-Britain	05/10/1985	100	0.69	A	1
2 Haiyuan, China	12/16/1920	220	12	B	12	39 W.Irian	11/17/1985	70	0.49	B	1
3 Kuyun	08/10/1931	180	8.5	B	12	40 Papua	02/08/1987	90	1.11	A	1
4 Parkfield, Ca.	06/07/1934	20	0.015	A	2	41 Alaska-I	11/17/1987	40	0.66	B	1
5 Turkey	12/26/1939	350	4.5	A	2	42 Alaska-II	11/30/1987	120	7.27	B	1
6 Imperial V. Ca.	05/19/1940	60	0.48	A	2	43 Alaska-III	03/06/1988	125	4.87	B	1
7 Turkey	12/20/1942	50	0.25	A	2	44 Burma-China	11/06/1988	85	0.37	A	1
8 Turkey	11/26/1943	265	2.6	A	2	45 Macqu. Ridge	05/23/1989	220	13.60	B	1
9 Turkey	02/01/1944	190	2.8	A	2	46 Loma Prieta, Ca.	10/18/1989	45	0.269	A	1
10 Darjung	11/18/1951	100	4.6	B	12	47 S.Fiji	03/03/1990	220	3.010	A	1
11 Turkey	03/18/1953	58	0.73	A	2	48 Sudan	05/20/1990	50	0.528	A	1
12 Alaska	07/10/1958	350	4.4	A	2	49 Philip.	06/14/1990	100	0.465	A	1
13 Gobi-Altai	12/04/1957	270	15.0	B	12	50 W. Iran	06/20/1990	150	1.350	A	1
14 N. Atlantic	08/03/1963	32	0.12	B	2	51 Philip.	07/15/1990	120	4.070	B	3
15 Aleutian	07/04/1966	35	0.23	B	2	52 Off-N.Calif.	08/17/1991	40	0.443	B	1
16 Gibbs f-z	02/13/1967	60	0.37	B	2	53 Turkey	03/13/1992	50	0.116	A	1
17 Turkey	07/22/1967	80	0.62	A	2	54 Vancouver-I.	04/06/1992	80	0.119	A	1
18 Borrego Mtn, Ca.	04/09/1968	37	0.11	A	2	55 Landers(CA)	06/28/1992	100	1.060	A	1
19 Iran	08/31/1968	95	0.67	A	2	56 Alaska-IV	08/07/1992	50	0.176	B	1
20 Sitka Alas.	07/30/1972	180	3.0	A	2	57 Kobe,Japan	01/16/1995	55	0.243	A	11
21 Luhuo	02/06/1973	110	1.8	A	2	58 W.Irian	03/19/1995	80	0.225	A	11
22 Yunnan	05/10/1974	45	0.065	A	2	59 Sakhalin	05/27/1995	70	0.432	A	11
23 Gibbs f-z	10/16/1974	75	0.45	B	2	60 Kashmir	11/19/1996	58	0.237	A	4
24 Atlantic	05/26/1975	80	7.0	B	7	61 Iran	05/10/1997	120	0.735	A	4
25 Guatemala	02/04/1976	250	2.6	A	2	62 Tibet	11/08/1997	170	2.23	A	13
26 Yunnan	05/29/1976	35	0.05	A	2	63 Balleny-Isl	03/25/1998	315	17.000	B	5
27 Tangshan	07/27/1976	140	1.8	A	2	64 Ceram	11/29/1998	90	4.48	B	4
28 W. Irian	09/12/1979	80	2.37	B	1	65 off-Taiwan	05/03/1998	60	1.83	B	4
29 Imperial V. Ca.	10/15/1979	50	0.07	A	1	66 Honduras	07/11/1999	32	0.122	A	4
30 Ca.-Mexico	06/09/1980	25	0.04	A	1	67 Izmit,Turkey	08/17/1999	140	2.880	A	6
31 Eureka, Ca.	11/08/1980	120	1.12	A	1	68 Hector Mines, Ca.	10/16/1999	45	0.598	A	10
32 Daofu	01/23/1981	46	0.13	A	2	69 Duzce,Turkey	11/12/1999	40	0.665	A	9
33 N.-Zealand	05/25/1981	100	5.00	B	2	70 S.Indian-Ocean	11/15/1999	35	0.330	B	4
34 Aegean-Sea	12/19/1981	54	0.23	B	3	71 Vanuatu	02/25/2000	80	0.507	B	4
35 Aegean-Sea	01/18/1982	50	0.09	B	1	72 Sulawesi	05/04/2000	70	2.44	B	4
36c Aegean-Sea	08/06/1983	40	0.12	B	3	73 S.Indian-Oc.	06/18/2000	105	7.91	B	11
37 Off-N.Calif.	09/10/1984	30	0.10	B	1	74 Tibet	11/14/2001	420	5.9	A	4

References: (1) *Pegler and Das* [1996]; (2) *Romanowicz* [1992]; (3) *Yoshida and Abe* [1992]; (4) NEIC catalog; (5) *Antolik et al.* [2000]; (6) *Delouis et al.* [1999]; (7) *Lynnes and Ruff* [1985]; (8) *Ruff et al.* [1989]; (9) *Akyüz et al.* [2000]; (10) *Kaverina et al.* [2001]; (11) *Henry and Das* [2001]; (12) *Molnar and Denq* [1984]. M_0 in 10^{20} Nm and L in km.

Three of these events were recently studied by *Henry and Das* [2001], and we used their length estimates. For the other 12, we obtained estimates of length based on the distribution of aftershocks of $M > 4$ in the month following the event, as given in the NEIC contribution to the Council of National Seismic Systems (CNSS) catalog [*Malone et al.*, 1996]. We only kept those events with a clearly delineated aftershock zone. We calibrated our procedure by comparing our estimates with PD96 for the subset of common events. In most cases, the bias in our catalog-based estimate is towards longer ruptures, as expected, and does not exceed on the order of 30km. In the process, we found significantly smaller L estimates than PD96 for several events in the time period 1980–1992. Two of these events occurred in the Aegean Sea in 1981–83. For these events, which stand out as outliers in the M_0 versus L plot for the PD96 dataset, we chose the catalog-based estimate. The fourth event occurred southwest of New Zealand in '81 in a region of poor station coverage, and was studied by *Ruff et al.* [1989], who estimated a much shorter length.

[10] The M_0 and L estimates for the combined dataset thus obtained are listed in Table 1. Most of the data follow the $n = 3$ trend, albeit with significant dispersion, except for the largest events (Figure 1). At the time of the *Romanowicz* [1992] study, only 4 data points were available for events of $M_0 > 1.5 \times 10^{20}$ Nm which did not fit that trend bracketed by $n = 3$ lines, and which she labelled "anomalous" (Alaska '87–'88 sequence and Macquarie '89). There are now 12 such events (including the North Atlantic 1975 event studied by *Lynnes and Ruff* [1985] which was not considered in *Romanowicz* [1992]). We note that 11 out of 12 of these large "anomalous" events occurred in an oceanic, often intraplate setting (except the 1990 Luzon, Philippine earthquake, as studied by *Yoshida and Abe* [1992]). We therefore separated our dataset into two subsets: subset *A* comprises mostly events that occurred in a continental setting, and/or which, if their moment is larger than 1×10^{20} Nm, follow the trend of events that occurred on the San Andreas and Anatolian faults, on which the analysis of *Romanowicz* [1992] was based. The second subset ("*B*") comprises the 12 large "anomalous" events mentioned above, four great earthquakes in central Asia, as well as smaller events occurring in an oceanic setting. The resulting separate M_0/L plots are shown in Figure 2. We infer that each data subset can be fit rather tightly with an $n = 1$ trend for the largest events. The change of scaling simply corresponds to a larger moment for events in subset *B* ($M_o \sim 5\times 10^{20}$ Nm) than for those in subset *A* ($M_0 \sim 0.8–1\times 10^{20}$ Nm). For both subsets, the change in scaling occurs for $L \sim 80$ km. For smaller events, the dispersion is large, but, on average, the best fitting $n = 3$ trend plots lower for subset *B*.

[11] This difference in the position of the break in scaling in each subset can originate either from a difference in W_o, or from a difference in $\Delta\sigma$. If we assume that W_o cannot be much larger for events that occur in oceanic versus continental crust (at most a factor of 2 difference), Figure 2 implies that subset *B* has larger $\Delta\sigma$ than subset *A*. In other words, in the latter case, the corresponding faults are weaker. This result is consistent with studies that have

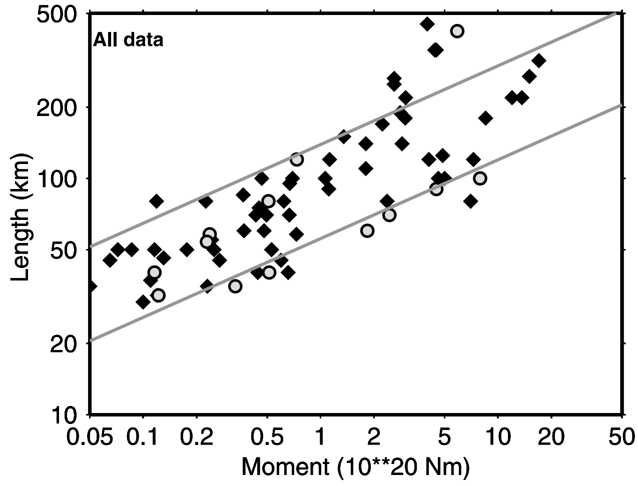

Figure 1. Moment-length plot for the dataset listed in Table 1. Lines corresponding to $n = 3$ bracketing most of the data have been drawn for reference. Circles correspond to recent data for which length was estimated from the NEIC catalog.

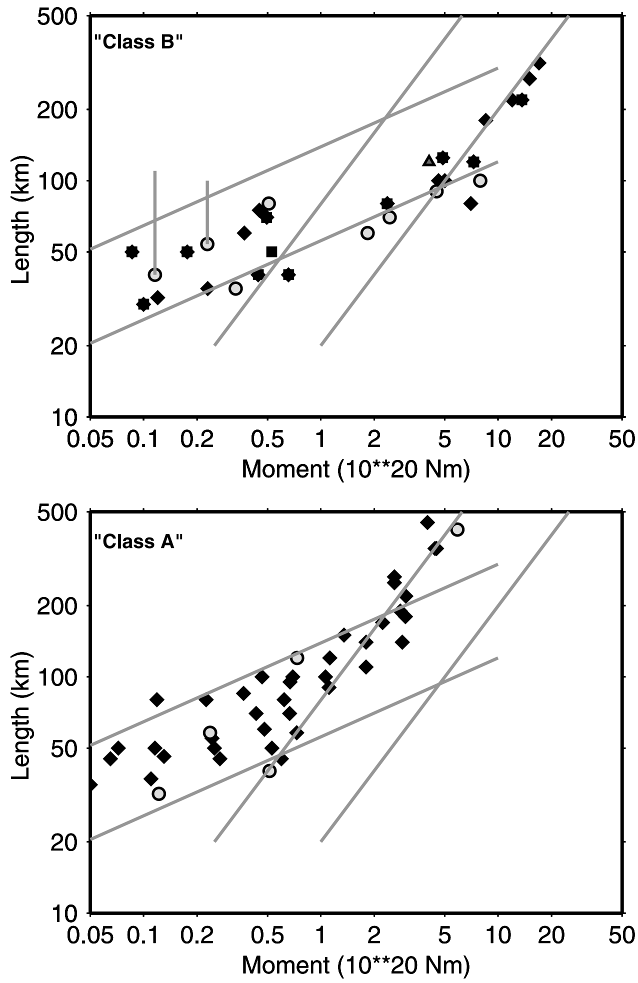

Figure 2. Moment-length plots for *A* (bottom) and *B* (top) events. Best fitting $n = 1$ trends are indicated for each subset of data. Circles as in Figure 1, diamonds from other sources (see Table 1). Triangle is Luzon '90 event. Vertical lines point to the length estimates of PD96 for Aegean Sea events discussed in text.

Table 2. Tests for Best Fitting Exponent in Moment-Length Scaling

Subsets of events	Minimum moment	num. of events	exponent	res. var.	n = 1	n = 2
Oceanic	5.0	7	1.17 ± 4.1	0.020	0.026	0.026
Oceanic	1.5	12	1.64 ± 3.9	0.045	0.044	0.059
Continental	1.0	16	1.09 ± 2.4	0.062	0.034	0.050
Continental	0.5	25	1.20 ± 1.4	0.137	0.048	0.092

compared intra-plate and inter-plate events [e.g. *Kanamori and Anderson*, 1975; *Scholz et al.*, 1986], or determined that a continental inter-plate fault such as the San Andreas Fault in California is "weak" [e.g. *Zoback et al.*, 1987]. Whether the distinction is intraplate/interplate or oceanic/continental is not clear, as there are exceptions one way or the other. The characterization simply by differences in the strength of faults is therefore more appropriate and may indicate that transform faults in young oceanic lithosphere, and major continental plate boundary strike-slip faults are weak, whereas strike-slip faults on "older" oceanic crust are stronger.

[12] If we allow for two classes of earthquakes based on strength of the corresponding faults, the M_0/L dataset can readily be explained in the framework of the "W-model", with a constant stress drop within each class. At the same time, this classification provides a way to identify which strike-slip faults are weaker or stronger, with a marked tendency of the global dataset to exhibit a bimodal distribution.

3. Discussion

[13] While we do not expect statistical arguments to be convincing in the presence of a relatively small dataset with large uncertainties, it is nevertheless interesting to examine how well we might be able to distinguish an $n = 1$ scaling from an $n = 2$ scaling, and how well we can determine the location of the break in scaling, when we consider the two subsets of large strike-slip earthquakes separately.

[14] We have performed two sets of experiments. In the first set of computations, we solve for the exponent n and the constant c in the relation $\log M_0 = n \log L + c$. In the second set of computations, we fix the exponent (i.e. $n = 1$ or $n = 2$), and invert for a and b in the equation $M_0 = aL^n + b$. In each experiment, residual variances are computed for 2 cases, depending where we choose to position the break in slope: at $M_0 > 4 \times 10^{20}$ Nm versus $M_0 > 1 \times 10^{20}$ Nm for *B* earthquakes, and at $M_0 > 1 \times 10^{20}$ Nm versus $M_0 > 0.5 \times 10^{20}$ Nm for *A* earthquakes. The results (Table 2) show that, based on variance reduction, it is not possible to distinguish an $n = 1$ from an $n = 2$ exponent for the subset of largest *B* earthquakes. However all other experiments favor an exponent equal (or close to) $n = 1$. The break in slope preferentially occurs for moments on the order of $4-5 \times 10^{20}$ Nm and 1×10^{20} Nm for *B* type and *A* type earthquakes, respectively. Assuming $W_0 \sim 20$ km, this implies a stress drop of 10–30 bars for *A* type earthquakes, and larger by a factor of 4–6 for *B* type earthquakes, which is consistent with other studies [e.g. *Scholz et al.*, 1986]. In reality, of course, complexities in fault zone structure will result in a more continuous, non-linear relationship between seismic moment and length, with no abrupt kink, but a more gradual change of trend for the largest earthquakes [e.g. *Miller*, 2002]. Clearly the "W-model" is an oversimplification. Also, different definitions of L may lead to different scaling [e.g. *Mai and Beroza*, 2000].

[15] The interpretation proposed here in terms of stress-drop differences is clear only for events with $M_0 > 0.5 \times 10^{20}$ Nm. For smaller earthquakes, even though the average stress-drop is higher for class *B* than for class *A* events, the dispersion in the data is very

large. This may be due to the proportionately larger variability in fault width and strength for smaller events.

4. Conclusions

[16] We conclude that the present global dataset of large strike-slip earthquakes is compatible with an n = 1 moment-length scaling, and that the scatter in the data can be largely explained by distinguishing two classes of events. Most continental inter-plate strike-slip earthquakes occur on weak faults, and most events on relatively old oceanic crust or in intraplate settings occur on stronger faults. *Kanamori and Allen* [1986] have related differences in stress drops for large earthquakes to their repeat times. We note that this is also compatible with the division that we have suggested: for example, the Alaskan '87–'88 sequence, the Balleny Islands '98 and the Macquarie Ridge '89 earthquakes are all considered rare events. It therefore follows that moment/length scaling for large strike-slip earthquakes is in agreement with the notion that earthquakes with longer repeat times occur on stronger faults, and result in larger moments than earthquakes with shorter repeat times, for the same length of rupture.

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