



[Back to the deformation and Stress Change Modeling home page](#)

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Two-way coupling between Vesuvius eruptions and southern Apennine earthquakes, Italy, by elastic stress transfer

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Abstract. During the past 1000 years, eruptions of Vesuvius have often been accompanied by large earthquakes in the Apennines 50-60 km to the northeast. Statistical investigations had shown that earthquakes often preceded eruptions, typically by less than a decade, but did not provide a physical explanation for the correlation. Here, we explore elastic stress interaction between earthquakes and eruptions under the hypothesis that small stress changes can promote events when the Apennine normal faults and the Vesuvius magma body are close to failure. We show that earthquakes can promote eruptions by compressing the magma body at depth and opening suitably oriented near-surface conduits. Voiding the magma body in turns brings these same normal faults closer to Coulomb failure, promoting earthquakes. Such a coupling is strongest if the magma reservoir is a dike oriented normal to the regional extension axis, parallel to the Apennines, and the near-surface conduits and fissures are oriented normal to the Apennines. This preferred orientation suggests that the eruptions issuing from such fissures should be most closely linked in time to Apennine earthquakes. Large Apennine earthquakes since 1400 are calculated to have transferred more stress to Vesuvius than all but the largest eruptions have transferred to Apennine faults, which may explain why earthquakes more commonly lead than follow eruptions. A two-way coupling may thus link earthquakes and Vesuvius eruptions along a 100-km-long set of faults. We test the statistical significance of the earthquake-eruption correlation in the two-way coupling zone, and find a correlation significant at the 95% confidence level.

Introduction

More than 700,000 people live on the flanks of Vesuvius volcano near Naples, and several hundred thousand live between Campobasso and Potenza, along the Apennine chain 50 km to the northeast ([Figure 1a](#)). An explosive eruption of Vesuvius today would likely claim 15,000-20,000 lives [*Scandone et al.*, 1993a], and the 1980 $M=6.9$ Irpinia earthquake in the southern Apennines killed 3000. A remarkable feature

of Vesuvius eruptions is that many coincide with southern Apennine earthquakes. *Bonasia et al.* [1985] and *Marzocchi et al.* [1993] found that large earthquakes and effusive eruptions are typically separated by 5-10 years, with more earthquakes preceding than following eruptions ([Figure 2](#)). *Marzocchi et al.* [1993] proposed that this coupling is statistically reliable, although they lacked a credible physical model to explain the correlation. The goal of this paper is to propose a physical model to interpret the coupling between eruptions at Vesuvius and normal-faulting earthquakes along the Apennines.

Examples of normal-faulting earthquakes closely associated with eruptions or volcanic inflation are seen worldwide. The largest eruption in the 1100-year history of Iceland occurred at the south end of the eastern volcanic zone in 1783. It was followed 15 months later by Iceland's largest historical earthquake, a $M=7.1$ event, 110 km to the west in the south Iceland seismic zone. Five Hekla eruptions were also accompanied by earthquake swarms in the south Iceland seismic zone, 70 km away [*Stefansson et al.*, 1993]. The 1975 $M=7.2$ Kalapana earthquake on the south flank of Kilauea volcano, Hawaii, was followed by dike intrusions and eruptions 10 km away in 1977 and 1983. These eruptions were in turn succeeded by a $M=6.1$ south flank earthquake in 1989 [*Dvorak*, 1994]. The 1959 $M=7.3$ Hebgen Lake, Montana, earthquake struck 50 km from the center of Yellowstone caldera, which inflated after and perhaps before the earthquake [*Holdahl and Dzurisin*, 1991]. Thus closely timed volcanic events within 100 km of large normal-faulting earthquakes are not rare and may share a common origin.

Several studies have appealed to static stress transfer as a possible triggering mechanism for either the earthquakes or eruptions in such event pairs. Volcanic inflation on the Izu peninsula of Japan was followed 1-3 years later by $M=6.7$ and $M=7.0$ events at distances of 15-30 km. Static stress calculations by *Thatcher and Savage* [1982] showed that, if the apparent fault friction was very low, Izu inflation increased the Coulomb failure stress on the two offshore strike-slip faults by about 0.1 MPa. During 1980-1986, six $M \geq 6.5$ earthquakes struck 10-40 km from Long Valley caldera, California [*Savage and Clark*, 1982; *Cockerham and Corbett*, 1987]; inflation may have preceded and accompanied the earthquakes and continues today. *Savage and Clark* [1982] calculated that three of the four faults that ruptured as the 1980 earthquake were brought 0.1-1.0 MPa closer to Coulomb failure by the inferred inflation in Long Valley. The 1978 eruption of the Asal rift, Djibouti, was followed by a series of earthquakes; *Jacques et al.* [1996] argued that stress transferred by the dike eruption triggered the earthquakes. Some 4 hours after the 1990 $M=7.7$ Luzon strike-slip earthquake on the Philippine fault, a $M=4.8$ shock struck 8 km from Pinatubo volcano 100 km away; 3 weeks later, rumbling was reported on the volcano; and 11 months later, Pinatubo erupted. *Bautista et al.* [1996] calculated that the static stress changes imparted by the Luzon earthquake increased the pressure by about 0.1 MPa in a magma conduit beneath Pinatubo, potentially promoting the eruption.

Because of the 2000-year record of Vesuvius eruptions, and 1000-year history of large Apennine earthquakes, the Vesuvius-southern Apennine region offers a unique natural laboratory to study earthquake-eruption coupling. The eruptive record is well documented following A.D. 1630 [*Carta et al.*, 1981], and the historical seismic catalogue is complete for $M > 5.5$ earthquakes since 1620 to present [*Tinti and Mulargia*, 1985]. Within this period we can exclude the possibility that the correlation might be an artifact of parallel reporting. In this paper, we consider the hypothesis that eruptions are promoted by changing the state of stress of a magma body, driving fluid and volatiles toward the surface, and that earthquakes are promoted by increasing the Coulomb stress on suitably loaded fault planes. Our goal is to quantify the stress changes associated with these historical events. The model provides a physical explanation for some of the historical southern Apennine-Vesuvius temporal correlation and may offer a useful approach for

investigating such occurrences elsewhere.

Observations

Southern Apennine Earthquakes

Geological and seismological investigations reveal that the largest central southern Apennine earthquakes of this century are events on normal fault planes striking parallel to the Apennine belt [Valensise *et al.*, 1993]. During the last four centuries, several $M > 6$ earthquakes occurred in the segmented seismogenic belt of the southern Apennines [Valensise and Pantosti, 1995; Valensise *et al.*, 1993; Boschi *et al.*, 1995b]. Earthquakes and the known or inferred fault segments are plotted in [Figures 1a](#) and [1b](#).

We compiled the historical and recent record of large earthquakes for which fault parameters have been proposed (Table 1). The peak intensities and magnitudes are from the Catalogo dei Forti Terremoti [Boschi *et al.*, 1995a], (hereinafter referred to as CFT1); seismic moments are from Boschi *et al.* [1997] (hereinafter referred to as CFT2) and G. Selvaggi (written communication, 1997).

Historical earthquakes for which little is known about the causative fault are listed in Table 2. This list includes the largest known event along the southern Apennines, the 1456 $M \sim 7.3$ earthquake ([Figure 1b](#)) [Figliuolo, 1989], which did not coincide with a Vesuvius eruption. Because of their proximity to Vesuvius, we include three shallow moderate events in 1796-1883 reported on Ischia island ([Figure 1a](#)), although they do not belong to the southern Apennines and are not used in the calculations.

Vesuvius Eruptions

The historical record of Vesuvius eruptions is marked by extreme variability. The documentation for eruption dates is fair from A.D. 79 and good from A.D. 1631. Scandone *et al.* [1993a] distinguish three types of Vesuvian activity: Large-scale explosive Plinian eruptions, whose eruptive volumes are larger than 1 km^3 ; intermediate sub-Plinian explosive eruptions, with eruptive volumes of $\sim 0.1 \text{ km}^3$; and small scale effusive eruptions with volumes of $\sim 0.01 \text{ km}^3$. Two Plinian eruptions occurred in 7900 B.P. ($\sim 2.4 \text{ km}^3$ [Rolandi *et al.*, 1993a]) and in 3800 B.P. ($\sim 3.0 \text{ km}^3$ [Rolandi *et al.*, 1993b]). The most recent Plinian event is the A.D. 79 Pompeii eruption of $\sim 3.0 \text{ km}^3$ [Arno *et al.*, 1987]. Sub-Plinian eruptions occurred during the interval A.D. 79 to 1631, including those in A.D. 472 ($\sim 0.4 \text{ km}^3$ [Rosi *et al.*, 1983]), 512, and 1139. The last sub-Plinian eruption occurred in 1631 with a volume of $0.2\text{-}1.1 \text{ km}^3$ [Scandone *et al.*, 1993b; Rolandi *et al.*, 1993c]; the last effusive eruption was in 1944 with 0.01 km^3 ([Figure 2](#)). When direct observations are absent, we use the volcanic explosivity index (VEI) assigned by Rosi *et al.* [1987] and Scandone *et al.* [1993b] to estimate the eruption volume. In Table 3 we list the final eruptions (FE) and intermediate eruptions (IE), for which $\text{VEI} > 2$, which occurred after 1631. A final eruption is a rapid emission of magma characterized by a fast lava flow, reaching the low outskirts of the volcano, usually accompanied by notable explosive phenomena; these events last a few days and are sometimes followed by a collapse of the crater [Scandone *et al.*, 1993b]. They are termed "final" because are followed by a period of repose. They are usually the most violent eruptions, even if there are some intermediate eruptions with comparable VEI.

Volcanologists distinguish between "open-conduit" conditions at Vesuvius, when eruptions are frequent but

generally not explosive and lava is often visible in the volcano's central cone, and "obstructed-conduit" periods when lava is generally not visible and eruptions are rare but highly explosive. The 1631 eruption heralded a change from obstructed to open-conduit conditions, which persisted through 1944. *Santacroce* [1983] argues that a continuous supply and storage of basaltic magma within a shallow magma reservoir (3-5 km depth) best explains the correlation between the length of the repose intervals, the volume of the eruptions, and the degree of evolution of the erupted magma. In this view, each of the ~27 eruptions during 1631-1944 temporarily drained the shallow magma body, which was then replenished over the succeeding years to decades. Once filled, the magma body and conduit would then become sensitive to the small but sudden stress changes we consider here.

Correlation between Eruptive and Seismic Activity

Bonasia et al. [1985] identified a 1000-year connection between southern Apennine earthquakes and Vesuvius eruptions ([Figure 2](#)), arguing that the link is most apparent during the past 300 years, when the record is most complete. They reported roughly simultaneous changes in cumulative seismic energy release in the Apennines and eruptive production of Vesuvius during 1688-1703, 1794-1805, and 1857-1861 and suggested that earthquakes generally preceded eruptions by 5-6 years. They also found that eruptions issuing from dikes or fissures oriented NE-SW (bold lines in [Figure 2](#)) most closely coincided with Apennine earthquakes. [Figure 2](#) shows that an episode of heightened eruptive activity of Vesuvius during A.D. 968-1139 was accompanied by Apennine earthquakes in 848, 990, and 1125. Neither earthquakes nor eruptions were common during 1139-1631, whereas both display unusually high activity since 1631. A series of Vesuvius eruptions occurred during 1631-1834, accompanied by $6.5 < M < 7.0$ Apennine earthquakes in 1688, 1694, 1732, and 1805. The largest eruptions of Vesuvius since A.D. 472 appear to be followed during the succeeding half century by smaller effusive eruptions and Apennine earthquakes. *Marzocchi et al.* [1993] performed a statistical cross correlation between the eruptions of Vesuvius [after *Carta et al.*, 1981] and southern Apennine earthquakes during 1631-1944 and found at the 90% confidence level that the major effusive-explosive eruptions tend to follow Apennine earthquakes by 6-13 years. Once we identify the mechanical zone of coupling, we will reanalyze the correlation in this paper.

Modeling Strategy

The twin hypotheses that motivate our analysis are that increasing the Coulomb stress on a fault promotes earthquake failure and that increasing the compressive stress in a magma body at depth and opening its near-surface conduits promotes volcanic eruptions. We perform calculations in an elastic half-space with uniform isotropic properties. We restrict our examination to the instantaneous elastic effects, although viscoelastic rebound of the lower crust or heated rock surrounding a magma body may modify the transfer of stress.

Although the stress changes modeled here are small relative to earthquake stress drops, there is growing evidence that small static stress changes promote normal-faulting earthquakes, volcano seismicity, and some eruptions. Triggering of earthquakes by small stress changes has been proposed as an explanation for normal-faulting sequences in the southern Apennines [*Nostro et al.*, 1997; *Troise et al.*, 1998], as well as in other extensional provinces [*Hodgkinson et al.*, 1996; *Hubert et al.*, 1996; *Caskey and Wesnousky*, 1997]. Such interaction is associated with stress changes of 0.01-0.30 MPa imparted to faults by nearby earthquakes. Still smaller stress changes influence the global occurrence of $M > 6$ normal-faulting earthquakes: *Tsuruoka et al.* [1995] found that normal events are strongly correlated with the solid earth and ocean loading tidal stresses, which have amplitudes of just ± 0.003 MPa. Eruptions and microearthquakes at

several volcanoes are also modulated or triggered by tidal stresses [Dzurisin, 1980; McNutt and Beavan, 1981; Rydelek et al., 1988], although it is unknown whether Vesuvius exhibits such a sensitivity.

Promoting Eruptions by Earthquakes

We calculate the change in stress on a magma body beneath Vesuvius caused by slip on Apennine normal faults. The faults considered are plotted in [Figure 1c](#). Each is assumed to undergo pure dip slip from the ground surface to a maximum depth dependent upon magnitude, and to dip 65° NE or SW (Table 4). Because the state of stress in the southern Apennines [Amato et al., 1995; Amato and Montone, 1997] and near Vesuvius [Bianco et al., 1998] is NE oriented extension ([Figure 1a](#)), it is most likely that dikes striking normal to the extension direction (e.g., parallel to the Apennine chain) form the magma conduits. To assess the effect of the stress change on a dike, we calculate the horizontal stress change normal to the dike, as shown in [Plate 1a](#).

However, we also consider other types of volcanic sources, such as a spherical magma chamber, a sill, or a dike oriented perpendicular to the Apennine chain. A well-mapped radial fissure system [Rosi et al., 1987] feeds magma to Vesuvius volcanic edifice. To evaluate the effect on a spherical magma chamber, we calculate the pressure change; for a sill, we examine the vertical stress change. [Plate 1b](#) shows the changes of stress caused by slip along the Apennine fault on a NE striking dike (horizontal stress change normal to the dike), and [Plate 1c](#) shows those on a spherical magma chamber (pressure changes).

Promoting Earthquakes by Eruptions

We calculate the Coulomb stress change on fault planes caused by voiding a buried magma body. Following King et al. [1994] and Nostro et al. [1997], the Coulomb stress change, DS_f , is

$$\Delta\sigma_f = \Delta\tau + \mu' \Delta\sigma_n$$

where $\Delta\tau$ is the fault shear stress change in the direction of slip, DS_n is the outward normal stress change on the fault, and μ' is the apparent friction coefficient. Thus the fault is brought closer to Coulomb failure when either the shear stress rises or the fault is "unclamped". We assign uniform elastic moduli appropriate for the crust (shear modulus of 3.0×10^4 MPa, Poisson's ratio of 1/4), and make most calculations with $\mu' = 0.4$, equivalent to laboratory values of friction and modest fluid pressure, but we also conduct a sensitivity test on μ' . Coulomb stress changes are resolved onto Apennine normal faults dipping 65° NE or SW ([Figure 1c](#) and Table 4), based principally on studies of the 1980 Irpinia earthquake [Pantosti and Valensise, 1990; Valensise et al., 1993].

Voiding of the magma body is represented by closing of the walls of a vertical dike, as shown in [Plate 2](#), collapse of an horizontal sill, or deflation of a spherical magma chamber. Appropriate parameters for the first two mechanisms are listed in Table 4. We use a point source to approximate a spherical cavity because the stresses imparted by point source are indistinguishable from those by a finite sphere at distances more than twice the depth of burial [McTigue, 1987], and the Apennine normal faults lie 5-10 source depths from Vesuvius. We neglect body forces arising from movement and redistribution of mass from the magma chamber to the volcanic edifice or the atmosphere.

Modeling Uncertainties

The principal shortcoming in this analysis is the unknown nature of the volcanic source and the depth of the magma body supplying Vesuvius. Several lines of evidence point to a shallow body: Petrologic analyses favor a source lying between 3 and 8 km depth [Bonasia *et al.*, 1985]. Microearthquakes are concentrated at 4-5 km depth and extend only to a depth of 6 km [Vilardo *et al.*, 1996]. Because the large observed gravity changes have not been accompanied by measurable surface deformation since 1959, mass is likely being added at a depth of about 2 km [Berrino *et al.*, 1993]. Nevertheless, two- and three-dimensional tomographic seismic imaging [De Natale *et al.*, 1998; Zollo *et al.*, 1997] has not identified a seismic low-velocity body in the uppermost 5-6 km beneath the volcano. Magma bodies may thus be shallow if restricted to small chambers or thin dikes, or a larger magma reservoir may lie below 6-10 km. Thus, although we regard a shallow NW striking dike or spherical chamber as the most likely magma sources, we also consider NE striking dikes and sills, at depths ranging from 2 to 15 km.

Apart from the magma body geometry, there are several key sources of uncertainty in the modeling. Neither the eruption volume nor the magnitudes of the historical earthquakes are well known, and their uncertainties undoubtedly grow with the antiquity of events. The location and the geometry of the faults which ruptured during historical events are constrained using macroseismic intensity data and geomorphological studies [Valensise *et al.*, 1993]. The model source faults we use are simpler than the Apennine faults; for example, rupture on an antithetical and several en echelon faults occurred in the 1980 Irpinia earthquake [Pantosti and Valensise, 1990].

Our analysis is restricted to static stress changes, and thus time-dependent effects are neglected. Viscoelastic stress relaxation below the dike after the eruption [Hofton *et al.*, 1995] or below the fault after the earthquake could extend the area of large stress perturbation gradually outward from the source with time. Discontinuities of the elastic modulus in the crust [Quarenì, 1990], as well as a change of the crustal thickness [Albarelo and Bonafede, 1990] between the coastal volcanic province and the Apennines, could also alter the transfer of stress. The time lag between event pairings, often measured in years or decades, suggests a time-dependent component of the stress transfer which could involve viscoelastic relaxation processes. Viscoelastic deformation can modify or enhance the transfer of stress and may be required to explain such time lags.

Owing to the distance range between Vesuvius and Apennine faults the uncertainties in fault geometry (strike and dip, length and width) and location do not strongly affect our modeling results. The assumption of normal-faulting for the historical earthquakes is well constrained by the uniformity of the regional stress geometry, which indicates a nearly horizontal extension NE-SW oriented and a vertical maximum principal stress [Amato *et al.*, 1995; Amato and Montone, 1997].

Results

Promoting Eruptions

The model results shown in [Plate 1](#) indicate that slip on the Apennine faults compresses a spherical magma chamber or a NW oriented (Apennine-parallel) dike and extends a NE oriented (normal to the chain) dike. The computed stress perturbation thus depends on the type of volcanic source.

A key finding of this study is that earthquakes in the 100-km-long section of the southern Apennine chain closest to Vesuvius can promote eruptions by compressing the magma reservoir at depth (either a NW striking dike or a spherical magma chamber) and extending some feeder conduits to the surface. No such

coupling exists, however, if the magma resides in a horizontal sill. The coupling is most efficient when magma is stored in Apennine-parallel dikes, and when feeder conduits or fissures are oriented normal to the Apennines ([Figure 3](#)). This is because if both reservoir and conduits were aligned parallel to the Apennines, the applied stress would diminish with depth, driving magma downward rather than toward the surface ([Plate 1a](#), section B-B'), but conduits oriented normal to the Apennines are slightly extended, enabling magma or volatiles to ascend ([Plate 1b](#)). This result is consistent with the observations of *Bonasia et al.* [1985] that eruptions issuing through NE oriented fractures are more closely linked in time with large Apennine earthquakes. The negative downward stress change gradient would also be reduced if the incompressibility of the shallow layers overlying the magma chamber were lower than the lower crust, owing to structural changes or hot fluid migration outward the magma chamber, as modeled by *Bonafede* [1995].

As a result of a $M=6.9$ Apennine earthquake in the position closest to Vesuvius (fault A in the [Figure 1c](#)), a dike striking parallel to the Apennines is compressed by 0.085 MPa at 5 km depth ([Plate 1a](#)); a spherical magma body at the same depth would be compressed by 0.030 MPa ([Plate 1c](#)). The compression grows with earthquake magnitude and the shallowness of the reservoir depth and is larger if the normal fault dips NE ([Figure 4](#)). The effect of a $M=6.9$ earthquake at several positions along the Apennine chain is shown in [Figure 5a](#).

Model fault A (site of the $M=6.8$ 1688 earthquake) and fault B ($M=6.6$ 1732 shock) are strongly coupled to Vesuvius, which is compressed by more than 0.05 MPa at 5 km depth. At the same depth, fault C ($M=6.9$ 1694 and $M=6.9$ 1980 shocks) and fault D ($M=6.8$ 1857 shock) produce a stress change lower by more than one order of magnitude. Thus static stress transfer would not promote Vesuvius eruptions if the earthquake is more than about 50 km from fault A along the Apennine chain, regardless of earthquake magnitude. Such uncoupled events include the $M=6.8$ 1857 shock to the SE or the $M=6.6$ 1915 event to the NW ([Figure 1a](#)), neither of which is associated with an eruption, in accord with such a model.

The $M=6.9$ 1980 Irpinia earthquake, a multisegment rupture centered on the position of fault C, is calculated to be poorly coupled to Vesuvius. The stresses imparted by the Irpinia earthquake are small and dependent on the magma body geometry ([Figure 5b](#)). Dikes oriented parallel to the Apennines are extended, dikes oriented normal to the chain are compressed, opposite to the effects of earthquakes located on faults A and B, and a spherical chamber is modestly compressed. Because the Irpinia earthquake produced neither an eruption nor a change in the rate of microearthquakes beneath Vesuvius [*Bonasia et al.*, 1985; *Vilardo et al.*, 1996], these observations are consistent with the coupling model, particularly if the magma reservoir is contained in a NW striking dike.

Promoting Earthquakes

A second key finding of our study is that voiding a magma chamber beneath Vesuvius can bring suitably located Apennine normal faults closer to Coulomb failure ([Plate 2](#)). As in the case of earthquakes promoting eruptions, the coupling is much more efficient for faults A and B (0.008-0.018 MPa of Coulomb stress change at 10 km depth) than for faults C and D, and is stronger for NE dipping normal faults and diminishes with the depth of the magma body ([Figure 6](#)). A 0.01 MPa Coulomb stress increase on a normal fault requires a 0.05 km^3 eruption from a 5-km-deep dike and a 0.08 km^3 eruption from a 15-km-deep dike ([Figure 7](#)).

Promotion of Apennine earthquakes by Vesuvius eruptions is more sensitive to magma chamber shape than

to its depth. Voiding a buried sill, for example, has no effect on Apennine normal faults ([Figure 8a](#)). Closing dikes oriented parallel to the Apennines most efficiently promotes Coulomb failure on faults A and B; opening a dike oriented normal to the chain weakly promotes Coulomb failure as well ([Figure 8a](#)). Thus an eruption that closes dikes oriented parallel to the Apennines and opens shallow fractures or fissures normal to the Apennines will most strongly promote earthquakes in the Apennines. This is the same geometry that favors stress transfer to the volcano from earthquakes. These results are only modestly sensitive to the apparent coefficient of friction on the fault, m' , with the Coulomb stress increasing with m' ([Figure 8b](#)).

Coupling Zone

Our modeling suggests that there exists a coupling zone where normal-faulting earthquakes may promote eruptions and vice versa. We thus define the "coupling zone" as the region comprising faults A and B, since the stress effects on Vesuvius caused by a normal faults located outside of the coupling zone and vice versa are one tenth than those caused by faults located inside it. This region is schematically represented in [Figure 1a](#) by a dashed line containing the closest faults to Vesuvius and bounded by the two fault zones of recent events in southern Apennines: The 1984 Lazio-Abruzzo earthquake (to the NW) and the 1980 Irpinia earthquake (to the SE).

The temporal patterns of the $M > 5.5$ southern Apennine earthquakes within the coupling zone and of the final eruptions of Vesuvius since 1631 is shown in [Figure 9](#). The temporal sequences of earthquakes and of the NE-SW fissure eruptions represent a selection of the two temporal sequences statistically analyzed by *Marzocchi et al.* [1993]. According to our physical model and previous observations [*Bonasia et al.*, 1985], we expect the two selected sequences to be better correlated than found previously for all sequences.

Statistical Analysis

The physical model proposed allows us to select from the catalogues of earthquakes and volcanic eruptions the events which should be more strictly coupled ([Figure 9](#)) and to test the cross correlation between these selected data. Note that this selection is independent from the subsequent statistical analysis.

The variables used in the correlation analysis are the number of earthquakes and the number of volcanic eruptions in fixed time windows t . The period investigated spans 1620 to the present, during which the catalogues are complete [*Tinti and Mulargia*, 1985; *Carta et al.*, 1981]; the two series are compared at a maximum time lag of 20 years. Since a false correlation may arise from the windowing effect, we verify the stability of the results by using $t=3$ and $t=5$ years. We consider eruptions associated with NW oriented near-surface conduits (as listed in Tables 1, 2, and 3). To verify the stability of the result, we perform our calculations when the first two eruptions (marked by b in Table 3) are either included (case A) or excluded (case B; [Figure 10](#)) from the analyzed data.

Because the two time series are composed by discrete values; we use a nonparametric statistical approach, the Spearman correlation coefficient r , that is efficient and produces unbiased results independently of the statistical distribution of the data [*Gibbons*, 1971; *Conover*, 1980]. Given a bivariate random sample of size N , $(X_1, Y_1), (X_2, Y_2), \dots, (X_N, Y_N)$, we let $R(X_i)$ be the rank of X_i as compared with the other X values, for $i=1,2,\dots,N$. $R(X_i)=1$ if X_i is the smallest of X_1, X_2, \dots, X_N , $R(X_i)=2$ is the second smallest, and so on, with rank

N being assigned to the largest of the X_i . The $R(Y_i)$ are chosen in the same manner. When we have ties, as in the present case, we assign to each tied value the average of the ranks that would have been assigned if there had been no ties.

The measure of the correlation is designated by r and, in the presence of ties, is defined as

$$\rho = \frac{E_{xy}}{\sqrt{E_x E_y}}$$

where

$$E_{xy} = \sum_{i=1}^n R(X_i)R(Y_i) - N\left(\frac{N+1}{2}\right)^2$$

$$E_x = \sum_{i=1}^n R(X_i)^2 - N\left(\frac{N+1}{2}\right)^2$$

$$E_y = \sum_{i=1}^n R(Y_i)^2 - N\left(\frac{N+1}{2}\right)^2$$

r is simply the parametric Pearson correlation coefficient computed on the ranks and average ranks [see, e.g., *Conover*, 1980]. A positive correlation means that the larger values of X (the number of eruptions) tend to be paired with the larger values of Y (the number of earthquakes), and the smaller values of X and Y also tend to be paired together; thus a high level of seismicity is associated with a high number of eruptions. If the larger values of X tend to be paired with the smaller values of Y and vice versa, then the measure of correlation should be negative. Negative time lags mean eruptive activity is followed by seismic activity; positive time lags mean eruptive activity is preceded by seismic activity.

An important issue for this analysis concerns the attribution of the statistical significance of r . The large number of ties contained in the two time series could compromise the estimate [see, e.g., *Conover*, 1980]. To overcome this problem, we assign the statistical significance at the largest value of correlation coefficients through a Monte Carlo simulation. We keep fixed the time series of the volcanic eruptions, and generate 1000 synthetic time series of earthquakes, shuffling the interevent times between adjacent seismic events. For each series we calculate the largest Spearman correlation coefficient at different time lags and compare it with the largest value calculated in the real case. The statistical significance, which represents the probability that one observes the largest value of r by chance, is estimated by $M/1000$, where M is the number of cases in which the synthetic series has the maximum value of r larger than the largest value observed.

The results of this statistical analysis for case B are reported in [Figure 10](#). The two largest peaks, calculated for $t=3$ and $t=5$ years, suggest earthquakes lead eruptions by 5-6 years, confirming the correlation between seismic activity and following volcanic eruptions found by *Marzocchi et al.* [1993]. Time-lags for case A are the same. We find a significance level of 0.04 for case A and $t=3$ years and $t=5$ years; a significance level of 0.01 for case B and $t=3$ years and 0.05 for case B and $t=5$ years. Thus the physical model proposed to explain the link between southern Apennine earthquakes and Vesuvius eruptions is consistent with the past 300-year record.

Conclusions

The boundary of the two-way coupling zone, within which normal-faulting events promote eruptions and eruptions promote earthquakes, is shown in [Figure 1a](#), and key examples of the stress field induced by these events are illustrated in [Plate 1](#) (earthquakes enhancing eruptions) and [Plate 2](#) (eruptions enhancing earthquakes). A 0.1 km³ eruption enhances the failure stress on Apennine faults about the same amount as a $M=6.4$ Apennine earthquake stresses a dike beneath Vesuvius. Because such sub-Plinian eruptions are rarer than $M=6.4$ Apennine earthquakes, one would expect earthquakes to more commonly precede than follow eruptions, consistent with our statistical findings.

Do static stress changes promote Vesuvius and Apennine events? The evidence is tantalizing but equivocal. We have provided a plausible mechanical mean by which such a coupling might exist by the transfer of 0.01-0.10 MPa of stress. The observation that some shield volcanoes and most $M>6$ normal-faulting earthquakes exhibit a tidal correlation (± 0.003 MPa) and that aftershocks of several large shocks are promoted or suppressed by stress changes of just ± 0.01 MPa means that it is indeed possible that the calculated stress changes are sufficient to trigger or enhance the occurrence of events. In addition, several internal consistency tests lend support to the inference: (1) The observed earthquake-eruption interaction is strongest within the mechanical coupling zone. (2) The magma geometry that most promotes stress transfer to the faults also most promotes eruptions by earthquakes. (3) Long periods are seen during which neither eruptions nor large Apennine shocks took place, such as A.D. 1150-1450. (4) The 1980 Irpinia earthquake transmitted little or no calculated stress to Vesuvius, in accord with the absence of an observable effect at Vesuvius.

We have reperformed the statistical analysis of *Marzocchi et al.* [1993] selecting earthquakes and eruptions according to the results of our static stress changes calculations. The significance is improved, suggesting that the physical model proposed here is consistent with the available observations. Despite these encouraging results, the historical data for Vesuvius are sparse and incomplete, and uncertainties about the magma reservoir and conduit geometry leave open the possibility that the stress coupling is weak.

The open-conduit condition that Vesuvius entered in 1631 may cause a stronger stress sensitivity than when the conduit is obstructed. During periods in which the conduit is open, the magma reservoir replenishes within 1-2 decades after the preceding eruption. After it is refilled, small stress or pressure changes would be much more effective in triggering subsequent activity. During periods when the conduit is obstructed, the stress required to trigger an eruption may be higher than the amount transferred by nearby earthquakes. This may explain why the largest earthquake in the coupling zone, the $M=7.3$ 1456 event ([Figure 1b](#)), did not precipitate an eruption.

Bonasia et al. [1985] proposed that strain episodes triggered by the earthquakes are connected to the volcano through a crustal block system, and *Marzocchi et al.* [1993] posited a "slowly propagating extensional strain pulse" from the normal faults, promoting magma generation or ascent. Other studies have appealed to extensional stress as the mechanism to trigger eruptions after earthquakes in subduction zones [*Barrientos*, 1994]. Our results indicate that increasing the compressive stress acting on a NW oriented dike and on a spherical magma chamber promotes eruptions, as well as the extension on a NE oriented (normal to the chain) dike, promote eruptions. Unlike previous studies, which examine only how earthquakes can promote eruptions, we also find that eruptions can bring normal fault closer to Coulomb failure, promoting earthquakes. Nevertheless, our physical model does not account for the observed time lags.

The chief limitation of studies such as ours is the insufficient knowledge of the geometry of the magma body, given the proximity of some earthquake-volcano systems: The closer the earthquakes to the

volcanoes, the stronger the stress transfer, but the more sensitive the results become to the detailed magma-fault geometry. At Vesuvius, vertical dikes result in the most efficient coupling, whereas horizontal sills exhibit none. Efforts now underway to image the magmatic system from three-dimensional seismic tomography should greatly limit the range of possible behavior and permit further tests of the stress transfer hypothesis. If successful, stress transfer could be applied to other sites with evident earthquake-eruption coupling, such as in California, Iceland, and Hawaii.

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FIGURES and TABLES

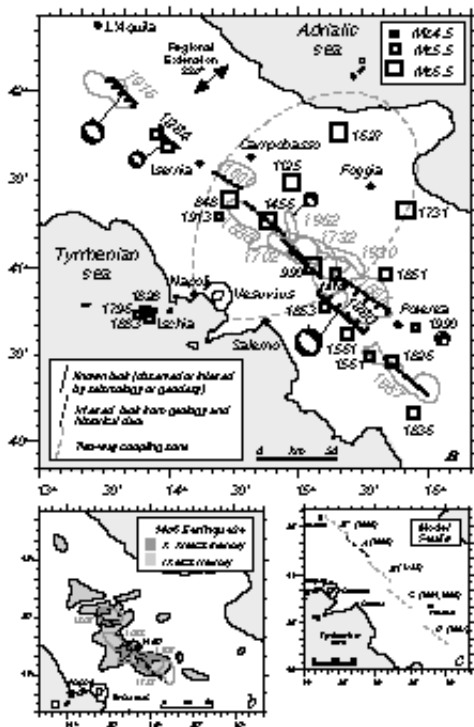


Figure 1. (a) Sites of historical earthquakes in south central Italy listed in Tables 1 and 2, associated faults, and Vesuvius volcano. Epicenters, or if unknown, the isoseismals of highest MCS degree (Mercalli-Cancani-Seeberg, termed "Modified Mercalli" in the United States), are shown. Focal mechanisms and borehole breakout data indicate a uniform extensional tectonic stress, whose minimum principal axis is horizontal and oriented nearly perpendicular to the Apennine chain [Hippolyte et al., 1994; Amato et al., 1995; Montone et al., 1995]. The "coupling zone" is the region in which normal-faulting earthquakes may promote Vesuvius eruptions and vice versa, according to our physical model results. (b) Isoseismals for the 1456 earthquake [Figliuolo, 1989] with 1688-1805 isoseismal zones. (c) Simplified model faults A to D used for calculations, with the dates of earthquakes that Valensise et al. [1993] associate with each fault segment. Fault B' is the same distance from A as fault B and thus experiences the same stress transfer from Vesuvius.

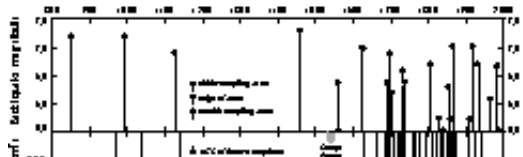
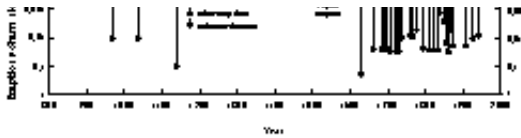


Figure 2. Earthquakes (lines extending upward proportional to magnitude, $M > 5.5$) during A.D. 800-2000 in the southern Apennines. The catalog is probably complete for $M > 7$ events after about A.D. 1400, and for $M > 5.5$ earthquakes after A.D. 1600



[Boschi et al., 1995a, 1997]. Eruptions (lines extending downward with line length proportional to eruption volume, $>0.001 \text{ km}^3$) are taken from Table 3.

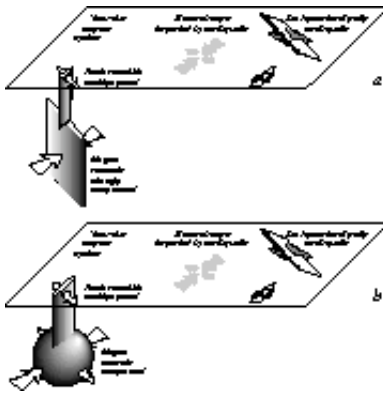


Figure 3. Schematic illustration of the response of a hypothetical Vesuvius magmatic system to a southern Apennine normal-faulting earthquake for (a) a buried dike in which at least one fissure or feeder conduit strikes NE and (b) a buried spherical magma chamber with a NE striking fissure.

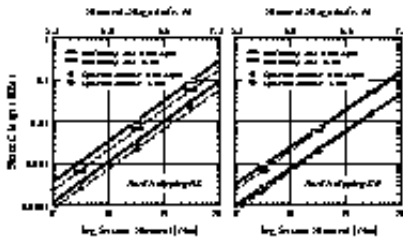


Figure 4. Stress change at Vesuvius caused by a normal earthquake at fault A, as a function of earthquake magnitude (or seismic moment), fault dip, and the depth and geometry of the magma reservoir. A $6.0 < M < 6.3$ Apennine shock exerts an effect equal to the diurnal tidal stresses (0.003 MPa); a $6.3 < M < 6.7$ shock transfers 0.01-0.02 MPa of stress, an amount found to be well correlated with aftershocks in other studies.

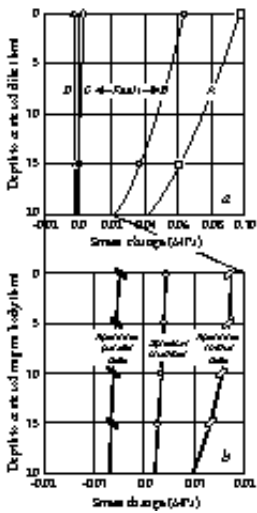


Figure 5. (a) Stress change at Vesuvius caused by a $M=6.9$ normal-faulting event at different sites along the southern Apennine chain, as shown in Figure 1c. In this example, faults dip NE, and the magma reservoir is assumed to be in a dike striking parallel to the Apennines. (b) Stress change at Vesuvius caused by the 1980 Irpinia earthquake, as a function of the geometry and depth of the magma reservoir. The Irpinia shock occurred close to model fault C but was a complex event on several fault segments. Magma geometries approximated by Apennine-parallel dikes and spherical magma chambers do not promote Vesuvius eruptions, consistent with the absence of eruptions or a change in the seismicity rate at Vesuvius in 1980.

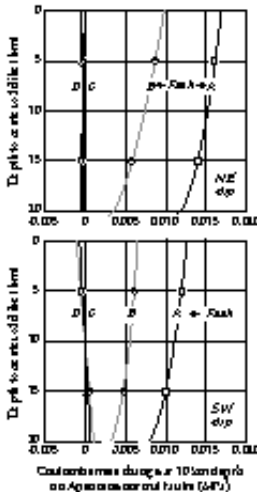


Figure 6. Effect of a 0.1 km^3 eruption issuing from a $5 \times 5 \text{ km}$ dike striking parallel to the Apennines on model faults A-D, as a function of the depth of the dike and the dip of the faults. Only faults A and B are brought closer to Coulomb failure, and the stresses are largest for shallow dikes and NE fault dips.

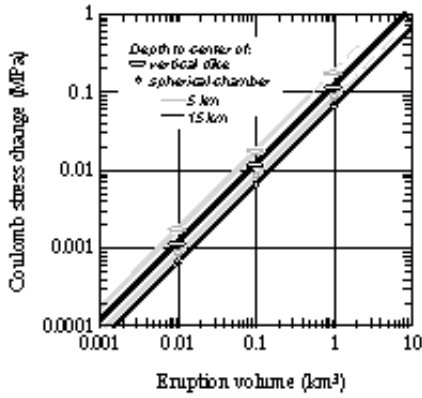


Figure 7. Coulomb stress change on fault A (dipping NE) caused by an eruption from a spherical magma chamber or 5×5 km Apennine-parallel dike, as a function of eruption volume and dike depth.

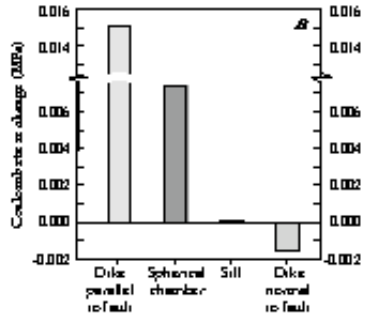


Figure 8. (a) Coulomb stress change on fault A (dipping NE) caused by a 0.1 km³ eruption from a 10-km-deep magma reservoir, as a function of the magma body geometry. The stress transfer hypothesis succeeds only if the magma reservoir lies in a Apennine-parallel dike or spherical chamber. (b) Coulomb stress change on fault A caused by a 0.1 km³ eruption from a 5-km-deep magma reservoir, as a function of the apparent friction coefficient and fault dip. Such an eruption brings faults closer to failure regardless of the friction coefficient.

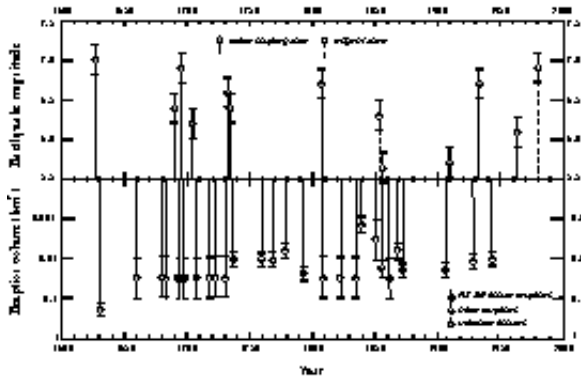
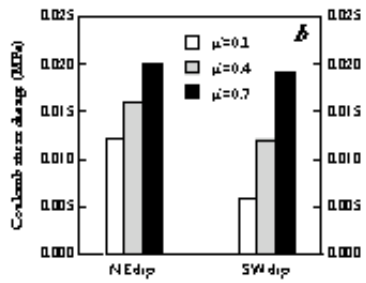


Figure 9. The temporal patterns of the southern Apennines $M > 5.5$ earthquakes (Tables 1 and 2) within the coupling zone (Figure 1a) and of the final eruptions of Vesuvius (Table 3) that occurred since 1600.

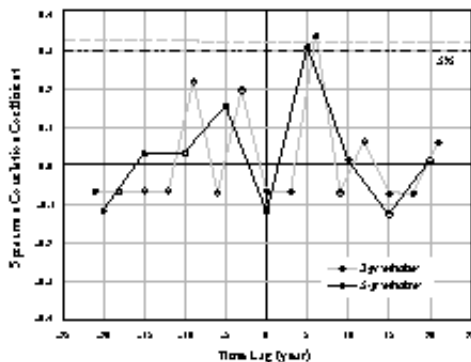


Figure 10. Spearman correlation coefficient for $t=3$ years and for $t=5$ years, as a function of the time lag, for case B: The first two eruptions (shown in Figure 9 and marked by *b* in Table 3) have been excluded from the selected data set.

Aug. 21, 1962 ^c	IX	6.1	5.9	3.2 ^d - 7 ^h	7	7	310, 65, 250 ^h	Benevento Sannio
Nov. 23, 1980 ⁱ	X	6.7	6.9	251 ^j	36 ^k	15 ^k	310, 60, 270 ^{b,k}	Irpinia
					10 ^k	10 ^k	124, 70, 270 ^{b,k}	
May 7, 1984	VIII	5.5	5.9	6.4 ^l - 7.8 ^m	7 ^l	7	123, 54, 264 ^l	Lazio-Abruzzo
May 11, 1984	VIII	5.4	5.5	2.0 ^m	5 ^l	7	170, 54, 293 ^l	Lazio-Abruzzo
May 5, 1990	VII- VIII	5.2 ⁿ	5.7	5.0 ⁿ	7	7	280, 79, 191 ⁿ	Potenza

a Magnitudes taken from CFT1 catalogue [*Boschi et al., 1995a*].

b *Boschi et al. [1995b]* and *Valensise et al. [1993]*.

c *M*?5.5 events located within the coupling zone.

d CFT2 catalogue [*Boschi et al., 1997*].

e *G. Selvaggi* (written communication, 1997).

f This study and *Nostro et al. [1997]*.

g *Ward and Valensise [1989]*.

h *Westaway [1987]*.

i Three subevents.

j *Giardini [1993]*.

k Top values refer to the first two subevents, while bottom values refer to the third subevent.

l *Westaway et al. [1989]*.

m *Rovelli et al. [1988]*.

n *Ekström [1994]* and *Azzara et al. [1993]*.

Table 2. Historical Earthquakes on Unknown Faults or With Moderate Magnitude

Date	I _{max} (MCS)	M ^a	Latitude ^b °N	Longitude ^b °E	Region
June 848	IX-X	7.2	41.40	14.48	Sannio
Oct. 25, 990	IX-X	7.2	41.03	15.10	Irpinia
Oct. 11, 1125	VIII-IX	6.9	41.48	14.95	Sannio-Molise
Dec. 5, 1456	XI	7.3	41.27	14.77	Southern Apennines
July 31, 1561	IX	5.5	40.63	15.38	Vallo di Diano
Aug. 19, 1561	X	6.4	40.50	15.55	Vallo di Diano
July 30, 1627 ^c	X	7.0	41.77	15.32	Gargano Promontory
March 20, 1731 ^c	IX	6.6	41.33	15.82	Foggia
March 18, 1796	VIII	5.5	40.75	13.92	Casamicciola (Ischia)
Feb. 1, 1826	IX	5.7	40.47	15.72	Basilicata
Feb. 2, 1828	IX	4.2	40.75	13.90	Casamicciola (Ischia)

Nov. 20, 1836	IX	5.5	40.17	15.88	Southern Basilicata
Aug. 14, 1851	X	6.3	40.97	15.67	Basilicata
April 9, 1853	IX	5.6	40.78	15.22	Irpinia
July 28, 1883	X	5.3	40.75	13.88	Casamicciola (Ischia)
June 7, 1910 ^c	IX	5.7	40.97	15.30	Irpinia
Oct. 4, 1913	VIII	5.2	41.29	14.38	Sannio

a Magnitudes taken from CFT1 catalogue [*Boschi et al., 1995a*].

b Coordinates are for the inferred epicenter (generally, the site of highest isoseismal intensity).

c $M > 5.5$ events located within the coupling zone.

Table 3. Final (FE) and Strongest Intermediate Eruptions (IE) of Vesuvius Volcano From 1631 to Present

Date	Type of Eruption	Type of Activity	Volume Estimate, ^a km ³	VEI
Dec. 15, 1631 ^b	FE	sub-Plinian	0.2 ^c -1.1 ^d	4
July 3, 1660 ^b	IE	explosive	(0.01-0.1)	3
March 26, 1680	FE	explosive	(0.01-0.1)	2
Aug. 12, 1682	FE	explosive	(0.01-0.1)	3
April 13, 1694 ^e	FE	effusive-explosive	(0.01-0.1)	3
May 19, 1698 ^e	FE	effusive-explosive	(0.01-0.1)	3
July 29, 1707 ^e	FE	effusive-explosive	(0.01-0.1)	3
June 6, 1717	IE	effusive-explosive	(0.01-0.1)	2-3
April 20, 1723	FE	effusive-explosive	(0.01-0.1)	3
Feb. 27, 1730	FE	effusive-explosive	(0.01-0.1)	2-3
May 14, 1737 ^e	FE	effusive-explosive	0.01	3+
Dec. 23, 1760	FE	effusive-explosive	~0.01	3
Oct. 19, 1767	FE	effusive-explosive	0.011	3
July 29, 1779	FE	effusive-explosive	0.0035	3+
June 16, 1794 ^e	FE	effusive-explosive	0.0235	3+
Aug. 12, 1805	IE	effusive-explosive	(0.01-0.1)	2-3
Oct. 21, 1822	FE	effusive-explosive	(0.01-0.1)	3+
Aug. 22, 1834	FE	effusive-explosive	(0.01-0.1)	3
Jan. 1, 1839	FE	effusive-explosive	0.0014 ^f	2-3
Feb. 5, 1850	FE	effusive-explosive	(0.001-0.01)	2
May 1, 1855	FE	effusive	0.017	2-3
Dec. 8, 1861 ^e	FE	explosive	(0.01-0.1)	3
Nov. 15, 1868	FE	effusive-explosive	0.006	2-3
April 24, 1872 ^e	FE	effusive-explosive	0.02	3
April 4, 1906 ^e	FE	effusive-explosive	0.02	3+
June 3, 1929	IE	effusive-explosive	0.012	2-3
March 18, 1944	FE	effusive-explosive	0.01	3

a Values in parentheses are estimated from the volcanic explosivity index, VEI.

b Eruption with unknown fissures.

c Scandone et al. [1993b]

d Rolandi et al. [1993c]

e Eruption occurring along NE-SW fracture system.

f Rosi et al. [1986]

Table 4a. Model Source Parameters for Faults

Fault \approx Event	Latitude, °N	Longitude, °E	Dip, deg	Strike, deg	Rake, deg
A \approx 1688	41.30	14.75	65	310	270
B \approx 1732	41.03	15.07	65	310	270
B' \approx 1805	41.52	14.45	65	310	270
C \approx 1980, 1694	40.73	15.35	65	310	270
D \approx 1857	40.37	15.85	65	310	270

Table 4b. Model Source Parameters for Earthquakes

M_w	M_0 , $\times 10^{18}$ Nm	L, km	W, km	Dip, deg	Strike, deg	Rake, deg	Upper Depth, km
5.6	0.32	9	9	65	310	270	0
6.3	3.1	14	14	65	310	270	0
6.9	28.8	32	15	65	310	270	0

Table 4c. Model Source Parameters for Dikes

Dike Orientation	Latitude, °N	Longitude, °E	Dip, deg	Strike, deg	L, km	W, km	Depth to Center, km
Apennine-parallel	40.821	14.426	90	310	5	5	5-15
Apennine-normal	40.821	14.426	90	220	5	5	5-15
Horizontal (sill)	40.821	14.426	0	310	5	5	5-15

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