SEISMICITY IN JAVA, INDONESIA: SPATIAL & TEMPORAL STUDY USING FRACTAL DIMENSION & B-VALUE

Research Proposal submitted to the Double Degree M.Sc. Program, Gadjah Mada University and Faculty of Geo-Information Science and Earth Observation, University of Twente in partial fulfillment of the requirement for the degree of Master of Science in Geo-Information for Spatial Planning and Risk Management

UGM

ITC

By :

AYU LINTANG RAMADHANI

10/309445/PMU/06818

AES 27671

Supervisor :

1. **PROF. DR. rer. nat. JUNUN SARTOHADI, M.SC. (UGM)**
2. **DR. DAVID G ROSSITER (ITC)**

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THESIS DRAFT

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Prepared By:
AYU LINTANG RAMADHANI
GMU: 10/309445/PMU/06818
ITC: 27671

Has been approved in Yogyakarta
On ……………………

Supervisor I, Examiner,

Prof. Dr. rer. nat. Junun Sartohadi, M.Sc. …………………….

Certified by:

Program Director of
Geo-Information for Spatial Planning and Risk Management

Prof. Dr. HA. Sudibyakto, M.S

Disclaimer

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ABSTRACT

The study about earthquake and seismic activity, from time to time has been developing. Despite of so many efforts to mitigate, to assess, to reduce the risk, or even to forecast earthquake, it still considered as a random and unpredictable events. Some research examined earthquake data in order to determine the similarity of the time and spatial pattern of seismic activity in tectonic zone. Self similar pattern which is the property of *fractal* geometry, can be measured, and later will be defined in a value of dimension. Box-counting and correlation integral are common used as the method of space covering. An area being observes its fractal dimension using sliding windows with different cell size. It determines the relationship between the numbers of non-empty cells of different cell size inside the region of space. At one fractal dimension may change with the epicenters cluster or fault location. Temporal change in fractal dimension, the time interval between two events of the same magnitude is approximately proportional to the magnitude of the event.

Another fractal property can be found in frequency-magnitude relation in Guttenberg-richter law. It estimated best using Maximum-likelihood.

Using data which heterogeneous, can caused temporary change in data frequency. To avoid this effect, it is necessary analyzing lower limit magnitude of an earthquake catalogue to get homogenous data set. The magnitude of completeness of the study area found to be 5,5 *Mw.* And the relation between fractal dimension and b-value are both positive and negative.

Keywords: fractal, b-value, box-counting, correlation integral, seismicity

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Ayu Lintang Ramadhani

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Abbreviations

BAPPENAS	National Development Planning Board
BMKG	Meteorology, Climatology and Geophysics Agency
BNPB	National Agency for Disaster Management
DJA	*Lembaga Meteorologi dan Geofisika, Jakarta, Indonesia* (Meteorology and Geophysics Agency)
EMSC	European-Mediterranean Seismological Centre
GS	USGS National Earthquake Information Service, Golden, Colorado, USA
HRV	Harvard University, Cambridge, Massachusetts
NEIC	National Earthquake Information Center
USGS	United States Geological Survey

1. INTRODUCTION

1.1 Background

Seismicity is a natural phenomenon which daily occurs. It caused a variety effects in earth surface. Some creates traceable ruptures and induces disaster, while others, being ignored because its powerless energy release. A powerful energy released, characterized first by increasing small scale seismic activity. Such a powerful energy release creates great earthquake and followed by accompanying effect.

The study about earthquake and seismic activity, from time to time has been developing. Despite of so many efforts to mitigate, to assess, to reduce the risk, or even to forecast earthquake, it still considered as a random and unpredictable events. A preliminary assessment which identifies whether there is a plate tectonics movement or not, is an example of an effort to shift signs into warning. Another common example is an analysis of earthquake data record which represents by numbers. Earthquake data record is an important part in seismic study. Some research examined earthquake data in order to determine the similarity of the time and spatial pattern of seismic activity in tectonic zone (Turcotte, 1989; Kagan, 1997; Bhattacharya, 2009). Such a statistical approach even if it is old-fashioned, gives good results to make a better understanding. One of statistical approach for seismicity which has been mentioned is finding the self similar pattern in the available data. Self similar pattern which is the property of *fractal* geometry, can be measured, and later will be defined in a value of dimension. *Fractal dimension* is used to represents correlation between complex spatial features in natural phenomena, such as earthquake. Application of fractal dimension in earthquake used to quantifies spatial clustering of events indicating seismicity of a region (Roy & Ram, 2006; Öncel & Wilson, 2002; Selvaraj et. al., 2010). A clustered earthquake event which is the natural behavior of strong earthquake events will be occurred, signed by decreasing value of fractal dimension itself.

Another fractal property can be found in frequency-magnitude relation in Guttenberg-richter law (Goltz, 1997). Some researcher tries to observe change in spatial and temporal distributions of frequency and magnitude of earthquakes (Öncel & Wilson, 2002; Roy et. al., 2010). It can represents ratio of small to large earthquakes within time period and estimates the expected magnitude in a time period. Relation of both, fractal and frequency-magnitude can be interpreted to understand the spatial and temporal relationship within the data (Öncel, 2002; Roy et. al, 2010).

1.2 Problem statements

Different fractal methods and estimations are used based on which aim to be achieved. Box-counting and correlation integral are common used as the method of space covering. Some researches deals with the observation of the fractal scale differentiation to see change across different spatial and correlate them with variables in environment for achieving a better understanding (Öncel & Wilson, 2002; Roy et. al., 2010; Mandal & Rodkin, 2011). For example, an area being observes its fractal dimension using sliding windows with different cell size. It determines the relationship between the numbers of non-empty cells of different cell size inside the region of space (Caneva & Smirnov, 2004). At one fractal dimension may change with the epicenters cluster or fault location. The relationship between those could be study, i.e. quantify and interpret it. As the fractal dimension changes with the changes in fault location, would be the interest to see changes in the relation of both. The fractal dimension also varies for different earthquake depth (Goltz, 1997; Bhattacharya et. al, 2010; Mandal & Rodkin, 2011). The fractal dimension of the worldwide epicenter and hypocenter pattern was found to be smaller for deep earthquake (280-700 km) than for intermediate earthquake (70-280 km) (Goltz, 1997).

While studying temporal change in fractal dimension, the time interval between two events of the same magnitude is approximately proportional to the magnitude of the event (Goltz, 1997). As this theory defined, fractal dimension can be a precursor for large earthquake events.

Fractal analysis has never been done in this study area, but there had been recent research about using fractal to map the earthquake events pattern in Indonesia. Its results conducted for the use of earthquake hazard zoning (Galih & Handayani, 2007).

Moreover the issue is not only to quantify the complex inter-relationship but also to manage and process data. Choosing the observed area from area of interest is an important key to analyze variation in spatial and temporal fractal dimension. By choosing it properly, the events from available earthquake data which not having impacts to the area of interest is eliminated. Above it all, indeed, a good and accurate earthquake data records required.

While in fact, identification of earthquake epicenter and its magnitude is varies, which caused by method diversity that used in capturing seismic waves. This lead different numbers and coordinate location in earthquake catalogue for each of agencies. Even in one earthquake catalogue, data record also varies. It caused by improving seismic record tools and methods. It is causing heterogeneity in earthquake data catalogue. Using data which heterogeneous, can caused temporary change in data frequency. To avoid this effect, it is necessary analyzing lower limit magnitude of an earthquake catalogue to get homogenous data set. Method that enables to be used is frequency-magnitude distribution by Guttenberg-Richter law which simplified as *b-value.*

There are many researches to develop methods estimating b-value. It aimed to get smaller standard deviations. Maximum likelihood and least square method are the common used estimation. This also became interest in this research because homogenous and more robust data set possible to produce which continue to analyze using fractal.

Comparing and relating both, positive or negative correlation of fractal and frequency-magnitude is related to the different modes of failure within the active fault complex, respectively the epicenter distribution (Goltz, 1997; Öncel & Wilson, 2002). For example, both b-value and fractal dimension rise significant, which indicate stress release occurs (could be connected with the process of disintegration of crustal rocks) through increased levels of low-magnitude and increasingly scattered seismicity (Öncel & Wilson, 2002; Mandal & Rodkin, 2011).

The current analysis aims at finding and analyzing the prominent features of the available data to represent the seismic distribution in geographical space and time of the study area. The formulation of issues to be tackled in this research is how and why to investigate spatial and temporal variations of fractal similarity pattern of epicenter distribution and frequency-magnitude distribution in study area from specific seismic catalogue.

1.3 Objectives

Objectives of this research is: "to investigate spatial and temporal variations of fractal similarity pattern of epicenter distribution and frequency-magnitude distribution in the study area from specific seismic catalogue"

1.3.1 Sub objectives

1. To investigate the lower limit magnitude of the earthquake catalogue of study area using different estimation
2. To calculate b-value of study area using different sliding windows accuracy
3. To calculate fractal dimension of study area using different sliding windows accuracy
4. To investigate the correlation between fractal dimension and b-value of study area

1.4 Research questions

The following research questions are expressed to keep this research to its purpose.

1. What is better estimation used for this catalogue? Why?
2. Which b-value estimation has the best visualization in mapping purpose? Why?
3. How does b-value change temporal and spatially in the study area?
4. How does fractal dimension change temporal and spatially in the study area?

5. What is the tendency of correlation between fractal dimension and b-value of the study area? Is it positive or negative? And what it indicates?

1.5 Benefit of the research

1. Contribution to hazard understanding

Through contour interpolation of fractal dimensions of epicenters and contour interpolation of b-value that will be generated in this research, earthquake will be more understandable, why such a high energy release occurs there. Finding similarity in earthquake events, and what makes them similar.

2. Methodological contribution

This research performed a statistical analysis that had not been used before to the study area. Methods that will use in this research can be applied into other hazards analysis. Methods also contributed to study the variation of spatial correlation with time.

2. LITERATURE REVIEW

2.1 Seismic hazard

Seismic hazard or Earthquake is “a term used to describe both sudden slip on a fault, and the resulting ground shaking and radiated seismic energy caused by the slip, or by volcanic or magmatic activity, or other sudden stress changes in the earth which affect activities of people” (www.earthquake.usgs.gov). From According to McGraw-Hill Dictionary of Scientific and Technical Terms, seismic hazard defined as “any physical phenomenon, such as ground shaking or ground failure, that is associated with an earthquake and that may produce adverse effects on human activities” (www.answers.com).

2.1.1 Faults & Faults System

When earthquake occurs, there are areas of stresses that left trace along the earth crust which is displacement of rock, called faults (Gates & Ritchie, 2007). Some faults are clearly visible on Earth’s surface. As examples of their evidence are displaced hills, offsets in rivers, and offsets in a shoreline across a fault. Other faults are difficult to observe because of their less evidence in Earth’s surface. Invisible faults may go undetected until their movement generates an earthquake. Determining a fault as active or inactive, also hard to be done, because it depends on how much harm the fault may cause if it should move significantly again and also considering the fact that not all active fault visible on the surface.

The fault’s length has been correlated with magnitude of earthquakes but it is not quite reliable for seismic potential because the difficulties to trace how much the rupture will extends. To better estimate the earthquakes magnitude, historical records is used. But then, another problem is that sometimes the historical record not extends in enough time to make reliable estimation.

Fault system consists of series of faults that interlinked each other.

2.1.2 Seismic waves

Seismic activity creates waves that can be felt in a range of miles away depends on its seismic energy and things that could distract the waves in the area. Its elastic waves may travel along earth surface or through inside the earth (www.earthquake.usgs.gov).

Seismic wave which moves along earth surface is *Surface wave.* And the one which arrived first on the seismic records and move through interior of the earth or through the volume of the planet called as *Body wave* (Gates & Ritchie, 2007).

Body waves are faster than surface waves, but in contrary, Surface waves are more powerful and destructive. There are two kinds of Body wave, P and S waves. Each type of wave shakes the ground in different ways. P waves shakes the ground back and forth in the same direction and the opposite direction as the wave is moving, while S waves perpendicular to the direction the wave is moving. S waves are the second arrive at the seismograph and travel only through the solid Earth, not liquid inner core of the Earth from the epicenter.

There are two kinds of Surface wave, Rayleigh and Love waves. Love waves travelling horizontally and perpendicular to the direction of travel. It side to side movements damaging structures and falling over sideways. Rayleigh waves having an elliptical motion only, and causing the ground shake like ocean waves. Energy of the waves decreases rapidly away from the epicenter and only damaging in the area of the epicenter.

2.1.3 Magnitude of earthquakes

To measure the energy released by an earthquake, as an indicator how big it is, Magnitude is used (Gates & Ritchie, 2007). Magnitude recorded by Seismograms in various methods. A type of seismogram is only using method that works over a limited range of magnitudes. There are methods which based on body waves, surface waves, and others based on different methods (www.earthquake.usgs.gov).

Types of magnitude based on USGS describe on Table 2.1.

Table 2.1 Types of Magnitude

Magnitude Type	Related Magnitude	Magnitude Scale	Distance Range	Description
Duration (M_d)		<4	0-400 km	Based on the duration of shaking as measured by the time decay of the amplitude of the seismogram. Often used to compute magnitude from seismograms with "clipped" waveforms due to limited dynamic recording range of analog instrumentation, which makes it impossible to measure peak amplitudes.
Local (M_L)	m_{bLg}	2-6	0-400 km	The original magnitude relationship defined by Richter and Gutenberg for local earthquakes in 1935. It is based on the maximum amplitude of a seismogram recorded on a Wood-Anderson torsion seismograph. Although these instruments are no longer widely in use, ML values are calculated using modern instrumentation with appropriate adjustments.
Surface wave (M_s)	M_D, M_R, M_Z, M_V, M_{GR}, M_{JMA}	5-8	20-180 degrees	A magnitude for distant earthquakes based on the amplitude of Rayleigh surface waves measured at a period near 20 sec.
Moment (M_w)	M_M,, M_E, M_t	>3.5	All	Based on the moment of the earthquake, which is equal to the rigidity of the earth times the average amount of slip on the fault times the amount of fault area that

				slipped.
Energy (M_e)		>3.5	All	Based on the amount of recorded seismic energy radiated by the earthquake.
Moment (M_i)		5-8	All	Based on the integral of the first few seconds of P wave on broadband instruments (Tsuboi method).
Body (M_b)	m_{bLg}	4-7	16-100 degrees (only deep earthquakes)	Based on the amplitude of P body-waves. This scale is most appropriate for deep-focus earthquakes. Represents the size of an earthquake from the very beginning rather than overall size.
Surface wave (M_{Lg})		5-8	All	A magnitude for distant earthquakes based on the amplitude of the Lg surface waves.

Source : www.earthquake.usgs.gov

The earthquake magnitude scale is one of the most fundamental earthquake source parameter to be used for catalogues. Different methods which used by seismograms to record earthquakes magnitude resulted different types of magnitude. Different types of magnitude, also means different scale. For example for surface wave magnitude, the scale is 5,6 M_s, while for moment magnitude, the scale is 4,9 M_w. Using uniform scale is desirable, to easily manipulate the data. How it can be done? See (section 5.1).

2.2 Fractal

2.2.1 Definition of fractal

In Oxford advanced learner dictionary, *Fractal* is “a curve or pattern that includes smaller curve or pattern which has exactly the same shape” (Hornby et. al, 2005). Goltz (1997) defined *Fractal* as invariant under geometric similarity (magnification) in every scaling transformation. Those geometric similarity means it is not use standard geometry or line segments. Invariant under geometric similarity which called as *self-similar* means it is stable or remains unchanged.

In contrary of those two definitions, fractal application for natural objects is not exact self-similar, but can be sensibly superimposed in statistical terms.

2.2.2 Fractal concepts of self-similarity

Self-similarity is one of the basic characteristic of fractal. Fractal in mathematical definition using infinite number of iteration, while in application of natural objects, it uses finite number of iteration. It is more to analytically self-similar in their statistical properties than empirically. This phenomenon is used to called statistical fractal (Goltz, 1997).

Natural objects such as coastlines, faults, and vegetations are the most often to be analyzed using fractal. For example, earthquake-fault has several fractal properties including distances and times between pairs of earthquakes, earthquake size distributions, and angles between focal mechanisms of earthquake pairs (Goltz, 1997; Roy, 2006).

2.2.3 Fractal in seismicity

Seismicity has fractal properties, as been mentioned in section 2.2.2. Fractal in seismicity applied based on its varies objectives. In earthquake understanding, Kagan & Knopoff, Mandelbrot said in their studies (as cited in Roy et.al., 2010), Fractal is a two-point spatial correlation function for earthquake epicenters which have a self-similarity characteristic of its finite iteration function. Fractal measured distance correlation of set of points (i.e. epicenters). Fractal analysis of seismic data drives spatial complexity into more spatial insight quantitatively.

In decades ago, the use of fractal in seismicity was tried to forecast future earthquake by calculating its epicenters in time series. It result that fractal in seismicity is more as a warning than predictions (Kagan, 1997). In present time, many application of fractal in seismicity used to measure the fault networks, epicenters, aftershocks to helps understand the cause of a high energy release in an inter-plate region and earthquake process in a range of scale micro to macro level (Roy & Ram, 2006). For example, fractal application in fault networks was done based on theory that earthquake generally occurs along faults. But in order to do that, accurate fault data or at least high quality of satellite imagery is required. If fault data not available, high quality of satellite imagery can be extracted to get its fault rupture as been done by (Gloaguen, 2007). In reality, not all faults can be detected its appearance on earth surface e.g. invisible faults in Yogyakarta regency which caused by volcanic ash sedimentation, as studied by Raharjo (as cited in Tsuji et al, 2009). Therefore, researchers still try to examine which feasible and proper fractal properties from available data for earthquake analysis. They use different methods and estimations. Indeed, those researches have one thing in common which values of fractal calculation represents as *dimension*.

2.2.4 Fractal dimension

Fractal dimension provides a quantitative measure of the spatial clustering of epicenters and the seismicity of a region for reflecting the character of organized structure and describes variability or shape of data for all possible scales (Goltz, 1997; Yuhua & Anjie, 2008).

Turcotte (as cited in Yuhua & Anjie, 2008) defined fractal dimension as:

$$N = \frac{C}{r^D} \qquad \text{...................................(1)}$$

Where r is the characteristic scale, e.g. time, length, coordinate, etc; N is the number of object related with value of r, e.g. output, predicted value, etc; C is constant, and D is the fractal dimension.

Fractal dimension has infinite numbers so that the method and details of estimation must be mentioned even type of fractal dimension is given. Otherwise, results are incomparable and could be meaningless (Goltz, 1997).

Method used to estimate fractal dimension i.e. ‘Ruler’ method, Box counting, Rescaled range (R/S), Roughness Length (RL), Variograms, Power spectrum, Correlation Integral, etc.

Types of fractal dimension are Eucledian dimension, Haussdorff dimension, Capacity dimension, Information dimension, Correlation dimension, and Generalized-dimension (multifractal).

A set of fractal dimension D possible to determine whether it is homogeneous by measuring its fractal dimension D_q (Roy & Ram, 2006). Fractal dimension D_q has values of q ranges from 2 to 22. The value of q = 0 known as the Capacity dimension (D_0) which use Box-counting method. The value q = 1 known as Information dimension (D_1). In two-dimension, values of q is 2 and it is known as Correlation dimension (D_2). For one fractal object, the value of these dimension is differ from each other. The generalized dimension D_q, defined for all real q with lower and upper dimension limit $D_{-\infty}$ and D_{∞}. Its limit related to the regions of the set (Roy & Ram, 2006). The lower bound determined by epicenter resolution, while the upper bound by the influence of the finite size of observed area, Kagan & Knopoff (as cited in Öncel, 2002).

In general, the common way to calculate fractal dimension for earthquake events are using capacity dimension and correlation dimension. Both are using the space covering techniques. More explanation about these methods explained in section 2.2.4.1 up to 2.2.4.2.

2.2.4.1 Capacity dimension

Capacity dimension or known as box-counting method is usually measures the spatial filling of events (in case of earthquake, could be set of epicenters or a fracture set) in a size of grid cells.

The set of points in three-dimension is covered with boxes of smaller size and the number of boxes which contain at least one point is counted at each resolution, as mentioned in Abarbanel et. al. (as cited in Goltz, 1997).

For example in two-dimension application, see Figure 2.1. The grids which contain at least one point are counted, while others are discarded. Different spatial resolution gives different result.

The general advantage of the box-counting algorithm is there are no boundary effects at all. This is due to the initial adjustment of the grid over the data.

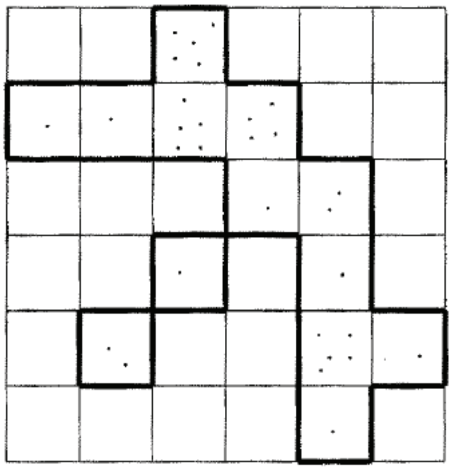


Figure 2.1 Box-counting methods (Goltz, 1997)

Self-similarity as the characteristic of fractal is can be found in frequency of occurrences of fault as a function of length on different scales. Faults which saw from centimeters to tens or even hundreds of kilometers are much the same. The areas of faults also follow the power law (Kulhanek, 2005).

Box counting method is given as follows:

$$N(r) \propto r^{-D_0} \qquad \text{.....................................(2)}$$

If an active fault system in an area is covered with a number of different sizes of square-boxes grids. Where the number of boxes $N(r)$ size r required covering the fault system is plotted as function of r on a double-logarithmic graph. D_0 estimated from the slope of the least-square regression line fitted to data between values $N(r)$ and r. A power-law relation exists between some linear distance r and the mass of a fractal (Goltz, 1997; Nanjo & Nagahama, 2004; Carranza, 2011).

For example, a research from (Öncel & Wilson, 2002) using Capacity Dimension D_0 to evaluate active fault networks. An observed area of a fault network is covered by square boxes with length r. And the number of boxes N which contain part of the fault pattern was counted as r was decreased. Its results increase of D_0 in observed area, which associated with denser and more complex regions of the active faults network. It also accommodates rupture on interconnected faults of larger total surface area and therefore larger seismic magnitude occurs.

From those research, it could inferred that Box-counting method mostly applied in surface phenomenon because it detects hierarchical set of cell sizes spatially, and gives great reliability and sensitivity of small changes (Nanjo & Nagahama, 2004; Carranza, 2011).

But if we examine on the circumstance in section 2.2.3, where sedimentary thickness give obstacles to directly observe the fault rupture from earth surface, the correlation dimension method is preferable. It provides a better and robust estimation of fractal dimensions of earthquake locations, Hirata; Kagan&Knopoff; Main (as cited in Mandal & Rodkin, 2011).

2.2.4.2 Correlation dimension

Correlation dimension D_c using correlation-function to measures clustering properties of a set of points (Grassberger & Proccacia, 1983). In practical, it is based on the distances between pairs of points of the set (i.e. epicenters) which determined by method of covering too. See Figure 2.2.

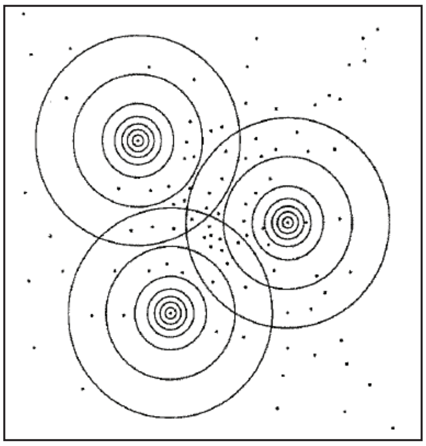


Figure 2.2 The correlation dimension (Goltz, 1997)

Figure 2.2 illustrate when the number of points within the varying seven radii of spheres around fixed three reference points is counted.

Harte (as cited in Kagan, 2007) mentioned several effects that may cause bias in the correlation dimension estimation. They are boundary effect, lacunarity (empty space or missing value), rounding effect, and noise or location error which only analyzed by doing simulations. Lacunarity of spatial earthquake distributions considered as a natural consequence of its statistical self-similar pattern.

Compared to the box-counting method which use hierarchical set of cell sizes so that a loss of information being compromised, method which used in the correlation dimension measurement is prefer to use to estimate fractal dimension for location of earthquake. However, box-counting method generally provides valid estimation for fractal dimension in surface system (i.e. faults), Okubo & Aki (as cited in Mandal & Rodkin, 2011).

The fractal spatial correlation dimension (D_c) provides a quantitative measure of the spatial clustering of events indicating the seismicity of a region (Öncel & Wilson, 2002; Roy & Ram, 2006). To measure the fractal dimension (D_c) of spatial distribution of seismicity is calculated from generalized correlation integral given by Grassberger & Procaccia (as cited in Selvaraj et. al. 2010):

$$D_c = \frac{\log C_q(r)}{\log r} \qquad \text{.....................................(3)}$$

where $C_q(r)$ is the correlation function, D_q exhibits a non-trivial scaling behavior for different values of q =1, 2, 3,…, and r is the length scale. For two-dimensional space measurements value of q is 2, while $C_q(r)$ defined below:

$$C_q(r) = \left\{ \frac{2}{N(N-1)} \sum_{j=1}^{N} \left[\sum_{i=j+1}^{N} H(r - |x_i - x_j|) \right] \right\} \qquad \text{.....................................(4)}$$

Therefore, $C_q(r)$ is proportional to the number of pairs of points of the fractal set separated by a distance less than r. Where N is the total number of pairs of vectors in the fractal set, H the Heaviside function; $H(z)=1, z \geq 0$ and $H(z)=0, z<0$. r_{ij} which is the distances between the points of a set which is obtained through spherical triangle method explained in eq. (5), and H the Heaviside step function.

Using eq. (3), the correlation fractal dimension for an integer q =2 estimate as an approach to two-dimensional fractal. In two dimensions, D_q, which from now on called as D_2, approaching a value of 2 signifies a uniform coverage of the plane. Uniform distribution of events that are self-similar is termed as mono fractal.

If the epicenters distribution has a fractal structure, this following relation should be obtained:

$$C_q(r) \propto r^{D_2} \qquad \text{.....................................(5)}$$

To obtain fractal dimension $D_{2,}$ the log-log plot of $C_q(r)$ and r should be done. Slope of the graph, will be fractal dimension $D_{2.}$

To estimate n degree between two given epicenters of latitude (degree) and longitude (degree), a spherical triangle is used. It assumed that epicenters are lie on the surface of a unit sphere. The distance between two epicenters using the Spherical triangle method defined by Hirata (as cited in Selvaraj et. al., 2010):

$$x_i - x_j = \cos^{-1}\left(\cos\varphi_i.\cos\varphi_j + \sin\varphi_i.\sin\varphi_j.\cos\left(\theta_i - \theta_j\right)\right) \qquad \text{........................(6)}$$

This calculation generates a linear function with slope D_2 as a fractal dimension of system.c Aki and Tosi (as cited in Roy et. al., 2010) said in their study, fractal correlation dimension value close to 0, means that epicenters clusters into one point; value close to 1, means that line sources are predominant; value close to 2, means that system is being filled up by a plane, and epicenters distributed randomly or homogenously distributed over 2-dimensional embedding space; value close to 3, earthquake fracture are filling up a volume of the crust. A high fractal correlation dimension where epicenter distribution less clustered, indicates that epicenter distribution can be concentrated along the barrier which gives the main shock (Roy, 2006).

2.3 b-value & magnitude of completeness

Spatial and temporal changes in frequency-magnitude distribution (b-value) can be used as an indication of stressed and relaxed patches in fault zones, Schorlemmer & Wiemer; Wiemer & Wyss (as cited in Mandal & Rodkin, 2011).

The statistical distribution of sizes for a group of earthquakes which is ratio of small to large earthquakes within time period simplified using *b-value* by Guttenberg-Richter law (as cited in Roy et.al., 2010):

$$\log_{10} N = a - bM \qquad \text{...................................(8)}$$

where N is number of earthquake, M is magnitude size and a is the total number of earthquakes. While b is constant generally takes a value close to one. Or, mathematically, b-value is the slope of the curve in the Guttenberg-Richter earthquake recurrence relationship.

The Guttenberg-Richter distribution applies to global catalogues of earthquakes or small regions. The Guttenberg-Richter distribution estimates how many earthquakes greater than or equal to the magnitude M can be expected in some time period for a given region, if accurate values of the parameters a and b are known. Therefore, the Guttenberg-Richter distribution plays a major part in earthquake forecasting and subsequent earthquake hazards modeling.

Common methods for estimating b-value are least square method, and maximum likelihood method. For number of events exceeding 50, the maximum likelihood method which is based on theoretical consideration is claimed to be a better robust and stable method than least-square method has been used to estimate b-value (Marzocchi & Sandri, 2003; Felzer, 2006).

$$b = \frac{\log_{10} e}{\overline{M} - M_c} \qquad \text{....................................(9)}$$

Where $\overline{M}$ is the average magnitude, and M_c is the threshold (cut off) magnitude which usually agree to the minimum magnitude for the completeness catalogue.

Earthquake data with magnitude below M_c often missed by network, because its magnitude is too small to be recorded on available stations, and because network operators eliminate events below a certain threshold because those are not of interest, and also in case of aftershocks, they are too small to be detected within the coda of larger events (Woessner & Wiemer, 2005).

b-value tends to decrease for smaller magnitude events due to the plot of logarithmic version of Guttenberg-Richter getting flatter at the low magnitude end of the plot. But in some case where the relation between logN and M supposed to be linearly, the frequency decreases more rapidly. Kulhanek (2005) describes some reasons for this problem. At small magnitudes, it is the incompleteness of data catalogs. And also because the small earthquakes are not as numerous as a constant b-value extrapolated from larger events so in a certain point, there is decline in the frequency. At large magnitudes, it is how the magnitudes are measured. It also because of the available data in catalogues is too short (with missing rarer large earthquakes). To solve this problems, it is better to use moment-magnitude (compared to body-wave and surface-wave magnitude) to indicate the size of large earthquakes.

Another problem in b-value calculation like underestimation in b-value needed to be considered. This caused by the use of earthquakes catalogue smaller than minimum magnitude completeness (M_c). The minimum magnitude of complete recording, M_c changes decreasingly with time in most catalogues, because the increasing number of seismographs and also the improving methods of analysis (Wiemer & Wyss, 2000). Therefore it is important to estimate the standard deviations of b-value. One of them is the formula provided by Aki, Shi & Bolt (as cited in Roy et. al., 2010).

$$\delta b = 2.3b^2 \sqrt{\frac{\sum_{i=1}^{N}\left(M_i - \overline{M}\right)}{N(N-1)}} \qquad \text{....................................(10)}$$

Where N is the number of earthquake and $\overline{M}$ is the average magnitude. Current estimation (eq.10) is used by Wiemer & Wyss as criteria to stabilize the "M_c by b-value Stability (MBS)" estimation by Cao & Gao for a better mapping purpose (Woessner & Wiemer, 2005).

Wiemer & Malone (2001) define a fast and reliable M_c estimation by using the point of the maximum curvature (MAXC) which is the maximum value of the first derivative of the frequency-magnitude curve (as cited in Woessner & Wiemer, 2005). The ease of this method, has a contradictive result, where M_c often underestimated for gradual curve of frequency-magnitude distribution that result from spatial or temporal heterogeneities (Woessner & Wiemer, 2005).

Woessner & Wiemer (2005) also modified an Entire-magnitude-range (EMR) method by Ogata & Katsura to estimate M_c. It uses a maximum likelihood estimator in two parts; to model the complete part, and to sample the incomplete part of the frequency-magnitude distribution. Again, to obtain a more robust estimation of magnitude of completeness (M_c) aimed at mapping case, entire magnitude range is used. It estimates M_c using entire data set of magnitude range, including the range of magnitude reported incompletely. Magnitude data which range above the assumed M_c, its b-value calculated using maximum likelihood method (Aki; Utsu as cited in Woessner & Wiemer, 2005). For below the assumed M_c, a normal cumulative distribution function $q(M \mid \mu, \sigma)$ which indicates the probability of seismic networks to detect an earthquake of a certain magnitude is fitted to the data.

$$q(M \mid \mu, \sigma) = \begin{cases} \dfrac{1}{\sigma\sqrt{2\pi}} \displaystyle\int_{-\infty}^{Mc} exp\left(-\dfrac{(M-\mu)^2}{2\sigma^2} \right) dM, M < Mc \\ 1, M \geq Mc \end{cases} \qquad \text{..........................(11)}$$

μ is the magnitude at which half of the earthquakes are detected. σ is the standard deviation which describe the width of the range where part of earthquakes detected. The sensitivity of detection capability in a specific network inversely related to the values of σ. The higher σ the more decrease detection capability of a specific network. μ and σ are estimated using maximum likelihood. To estimate log likelihood function for μ, σ, a, and b, the best fitting model with cumulative distribution function is used.

Mogi and Wyss in their study conclude that lower b-value corresponds to homogeneity in rock material which tends to high stress and shorter recurrence time, vice versa (as cited in Kulhanek, 2005).

2.4 Fractal dimension *D* and b-value relation

The b-value characterizes the fractal dimension in the possible values of energy of earthquakes, in comparison, *D* provides a measure of fractal dimension of earthquake location spatially, Aki; Turcotte (as cited in Mandal & Rodkin, 2011).

Caneva & Smirnov (2004) concludes that the decreasing of *b* value will be a sign for a large earthquake/strong earthquake (magnitude), vice versa. And the decreasing of fractal dimension is a sign of a clustered earthquake event which is the natural behavior of strong earthquake events will be occurred.

The spatial fractal dimension *D* of earthquakes is often correlated with the slope b of the Gutenberg–Richter law, independently of earthquake size. The relation between the fractal dimension *D* and the *b*-value of the Gutenberg–Richter law depends on earthquake magnitudes. With Fractal dimension *D* of small to large earthquakes events, results the distribution of small earthquakes/small faults within volumes, while large earthquakes/large faults along lines (Legrand, 2002). For example, Öncel & Wilson (2002) have studied spatial and temporal change of b-value and fractal dimension of two different study areas. In case of Japan, b-value and D_0 has positive variations which identify that more intensely fault areas accommodate stress release on smaller fault strands. In case of Northern Anatolian Fault Zone, it found to be non-stationary correlations between b-value and fractal dimension. It suggests that stress release occurs through increased levels of low-magnitude and increasingly scattered seismicity.

Other example given by Caneva & Smirnov (2004), which shows variations of the b-value and fractal dimension in Colombia as a function of depth. Around 100 km depth, the b-value tends to increase, followed by decreasing fractal dimension. It suggests that the counter-phase variation with depth around 100 km is related to the pressure and temperature conditions at those depths.

Comparing and relating both, positive or negative correlation of fractal and frequency-magnitude is related to the different modes of failure within the active fault complex, respectively the epicenter distribution (Goltz, 1997; Öncel & Wilson, 2002). For example, both b-value and fractal dimension rise significant, which indicate stress release occurs (could be connected with the process of disintegration of crustal rocks) through increased levels of low-magnitude and increasingly scattered seismicity (Öncel & Wilson, 2002; Mandal & Rodkin, 2011).

2.5 Research of Fractal analysis of seismicity in Indonesia

Galih & Handayani (2007) generates an earthquake hazard area zoning by differentiating each region of its earthquake hazard risk level using fractal analysis. USGS seismic catalogue grouped into 7 groups; Sumatera, Kalimantan, Sulawesi, Papua, Bali, South east Nusa and surroundings, Maluku-Halmahera-Banda and surroundings. Each of regions being calculated its magnitude-frequency relations using Guttenberg-Richter formula. Its b-value generated from Guttenberg-Richter formula, being used as an input for Fractal formula of Turcotte (1989):

$$D = 2 * b \qquad \text{....................................(12)}$$

The formula results fractal dimension D from each of region in Indonesia. It is value of fractal dimension D for each region which then used as an earthquake hazard area zoning. Maluku-Halmahera-Banda and surroundings has the highest fractal dimension, which is 1,53. Java has fractal dimension of 1,05. While Kalimantan has the lowest value of fractal dimension, which is 0,62. Galih & Handayani (2007) concluded that high number of earthquake events does not bring a region having high level of earthquake vulnerability. And fractal dimension is related to the potential earthquake events.

3 STUDY AREA

3.1 Location of the study area

The study area range between whole Java island and its surrounding sea, which lies in 5 - 12 S and 105 – 115 E. Java island is located in the south of Indonesia.

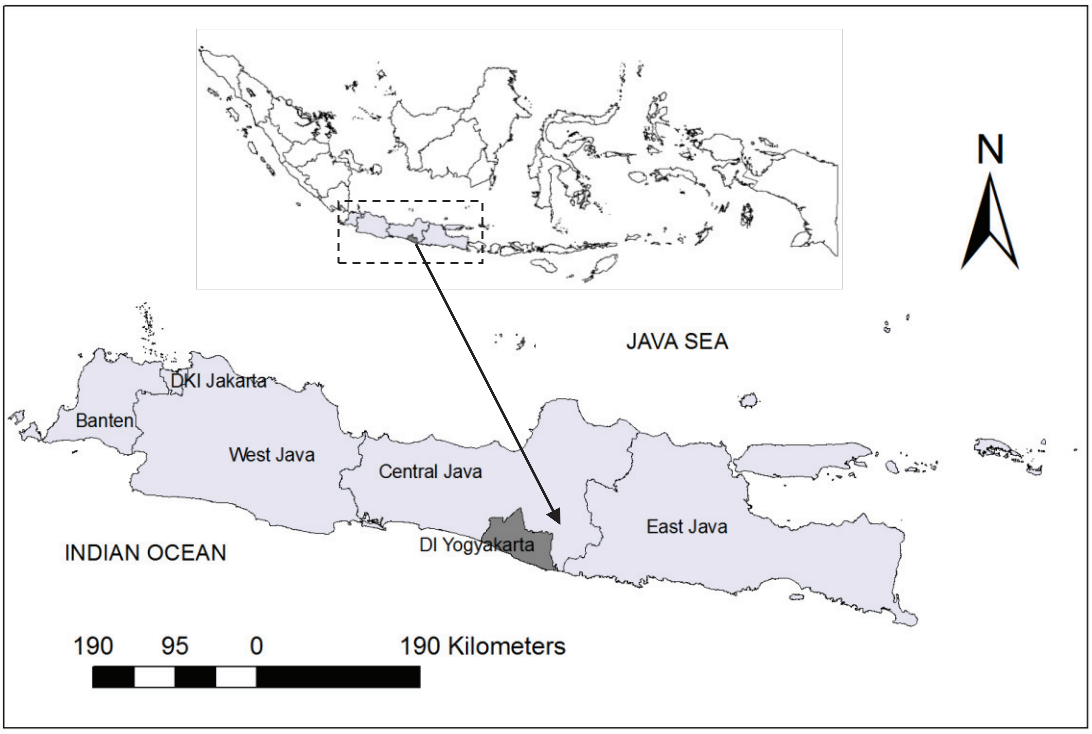


Figure 3.1 Location of Java island

3.2 Tectonic & geological settings

According to Mid-America Earthquake Center (2006), earthquakes around Java island show two distinct features which are earthquakes to the north arc of deep focus, whilst those to the south have shallower origins of nucleation. It may be as the impact of the edge of the overriding plate (the Sunda microplate) undergoing deformation due to subduction friction which caused intraplate earthquake on different existing faults due to bending and other stresses. See Fig.3.2.

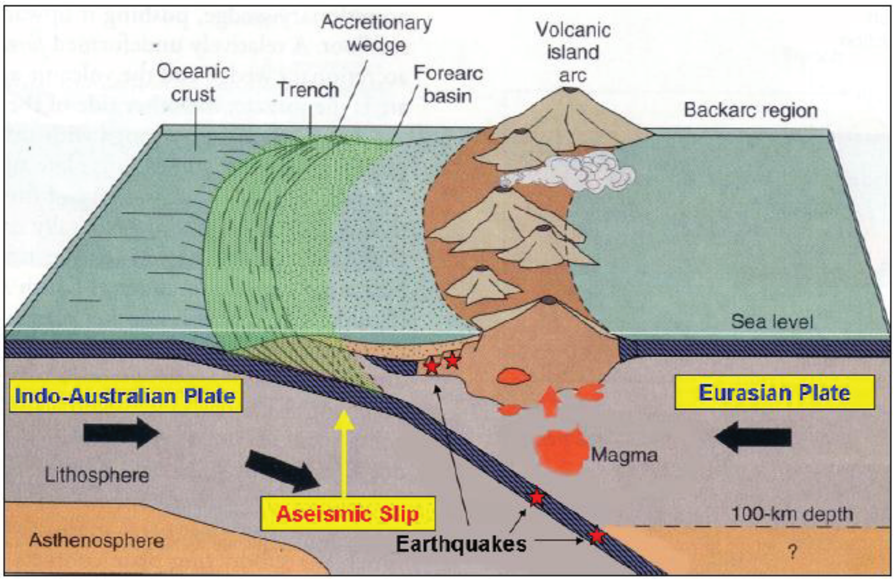


Figure 3.2 Interplate and intraplate earthquake potential in Java (Mid-America Earthquake Center, 2006)

Java close to the subduction zones of Indo-Australian plate and Eurasian plate, studies by Hamilton (as cited in Daryono, 2010). Moreover, the pressure of Indo-Australian plates in south of Java island drive a fault structure. The movement of those Indo-Australian and Eurasian plates, plus the activity of faults in land conducts Java to become a region with high seismic activity. One of large earthquake events which occurred in Java is 2006 Yogyakarta earthquake.

The geological conditions which commons in volcanic island contain layers of shallow, unconsolidated sediments. In Yogyakarta, those are a result of erosion processes and Merapi volcano eruption material. Young sediment around Bantul and Yogyakarta basin overlie more consolidated-rock. The seismic produced by the May 27 2006, earthquake encounters this low velocity, near surface sedimentary layer: the amplitude of the wave increases, the wave is bent toward vertical, and the wave becomes trapped in the near surface layer. This wave amplification produced very intensive damage around Bantul.

3.3 Historical seismicity

Seismic distribution in Java and its surroundings sea based on magnitude range is shown in Fig 3.3.

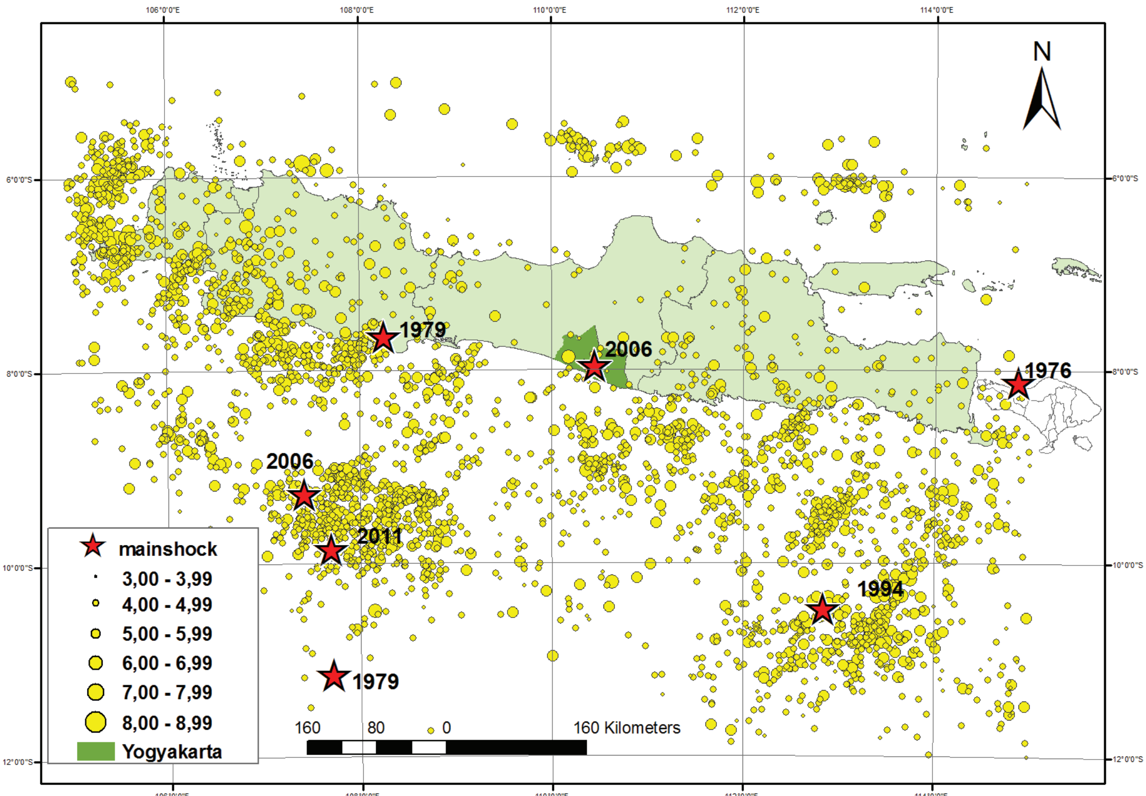


Figure 3.3 Seismic distribution in Java island and its surrounding sea (USGS)

According to USGS seismic data catalogue, seismicity mostly occur in south of Java island, Indian ocean. Most of them were events with medium up to strong magnitudes. Red stars indicate some of strong earthquake events in observed area.

4 METHODOLOGY

4.1 Data requirements

All seismic data gathered from secondary source. For seismic data catalog must be at least comparable and possibly larger than the return period of the largest expected event. Those seismic data are:

4.1.1 Global catalogue : USGS

One of global catalogue with ease of access and good quality of data is provided by USGS. USGS has several database of seismic catalogue; USGS/NEIC, USHIS for significant U.S. earthquakes, CDMG for California year 1735-1974, EPB for Canada year 1568-1992, India year 1063-1984, NGDC for Mexico-Central America-Carribean year 1900-1979, SISRA for South America year 1471-1981, SRA for Eastern-Central-Mountain States of U.S. year 1534-1986. This research is using USGS/NEIC database which can be downloaded from http://earthquake.usgs.gov/earthquakes/eqarchives/epic/epic_rect.php. Input of data expands up to the whole Java island, which its coordinate is 5 - 12 S and 105 – 115 E. The reasons why this research using the earthquake catalogue for all Java and its surrounding sea are: (1) earthquakes which epicenter are not in the province still cause damage in the study area, (2) limiting the earthquake catalogue into the local epicenters is not satisfied the requirements of good quality seismic data record in order to get the b-value estimation.

From USGS/NEIC data catalogue, sample of data obtained is shown in Table 4.1.

Table 4.1 Sample of USGS seismic data catalogue

Year	Month	Day	Lat.	Long.	Depth (Km)	Magnitude	Magnitude Type
1973	1	22	-7.57	107.26	88	4.9	*mbGS*
1973	2	1	-6.53	106.96	77	4.5	*mbGS*
1973	3	12	-9.41	111.14	38	5.4	*mbGS*
1973	3	24	-9.16	111.10	33	4.9	*mbGS*
1973	3	30	-8.71	112.59	97	4.7	*mbGS*
1973	7	22	-6.67	105.61	75	5.4	*mbGS*
1973	7	27	-8.97	106.93	61	5.5	*mbGS*

1973	10	14	-8.89	110.73	70	4.9	*mbGS*
1973	11	26	-6.76	106.59	62	4.9	*mbGS*
1973	12	28	-11.66	114.72	28	4.9	*mbGS*
1974	2	24	-6.36	105.08	93	4.9	*mbGS*
1988	10	9	-9.67	108.75	25	5.2	*MsGS*
1989	8	4	-6.84	106.13	33	5.2	*MsGS*
1989	9	12	-9.02	110.50	33	5.3	*MsGS*
1990	1	5	-8.80	106.44	29	5.8	*MsGS*
1994	6	4	-10.79	113.34	30	5.2	*MwHRV*
2001	6	13	-6.89	109.07	226	4.5	*MDDJA*
2002	2	10	-8.66	112.55	92	4.6	*MDDJA*
2002	3	6	-9.20	112.36	33	5.2	*MLDJA*
2002	4	3	-10.21	113.38	33	4.9	*MDDJA*
2002	4	14	-8.84	114.98	33	4.0	*MLDJA*

Source : Data processing, 2011

The total seismic data of Java island within coordinates 5 - 12 S and 105 – 115 E is 3158 records. As shown in Table 5.1, the type of magnitude is varies. It should be homogenized first.

The abbreviations after magnitude type i.e. mb(GS), Ms(GS), Mw(HRV), MD(DJA), ML(DJA) is related to the seismic stations which record those earthquake events (www.neic.usgs.gov). See Abbreviations.

4.2 Tools & equipment

Tools & equipment being used in this research described in Table 4.2, includes type of research tools & equipment, data, its function in this research, and availability status.

Table 4.2 Tools & equipment

Tools & equipment	Function
Seismic data record of study area	Main data
Microsoft word 2007	Words processing
Microsoft excel 2007	Numbers and graph processing device
ArcGIS 9.3	Spatial data analysis & visualization

Matlab 6.5.1	Running the Z-map software
Zmap v.6	Seismic data analysis & visualization

4.2.1 Zmap v.6

Zmap v.6 is a open-source software package to analyze seismicity. It constructed using Matlab sotware; a package widely used among researchers in the natural sciences. Zmap v.6 is a graphical user interface (GUI) - based software. It designed to help evaluate the catalogue quality and addressing specific hypotheses using several tools. See Table 4.3.

Table 4.3 Zmap v.6 tools & functions

Tools	Function
Histograms	Histograms of magnitude, depth, time,hour of the day
Data Import	Data import as ASCII, column separated files, using one of several existing input format filters or a custom-designed one.
Time Series Analysis	Cumulative number of events, time-depth plots, time-magnitude plots, cumulative moment release. Significance of rate changes using z, ß, and translation into probability.
Data Subset Selection	Select data inside or outside polygons, cut in magnitude, depths, or time.
Maps	Maps of seismicity; legend by time, depth, or magnitude. 3D view and rotation hypocenters. Cross-sections with one or multiple segments.
Frequency-magnitude Distribution	Estimating *a* and *b* values and uncertainties using maximum likelihood or weighted least squares as a function of depth, time, and magnitude. Map *b* and *a* values in map view, cross-section, or 3D. Compute local recurrence time maps. Differential *b* value maps for two periods. Create synthetic catalog with constant *b*.
Magnitude of Completeness	Estimate magnitude of completeness based on the deviation of the FMD from a power law. Analyze *Mc*

	as a function of time or depth. Map *Mc* in map view or cross-section.
Fractal Dimension	Compute the fractal dimension of hypocenters based on the correlation integral. Create maps and cross-sections of the fractal dimension.

Source: Wiemer & Malone, 2001

4.3 Research framework

The methodology of this research illustrated in the following Figure 4.1:

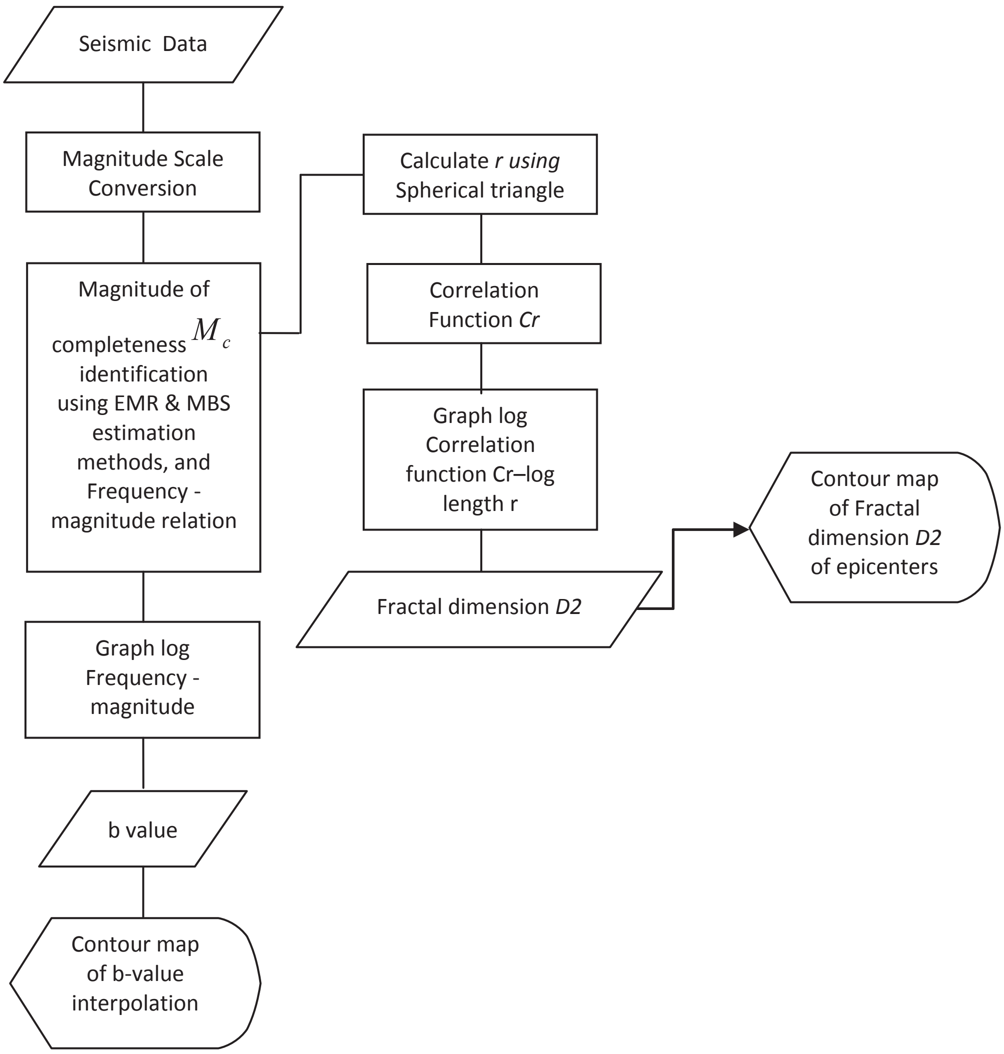


Figure 4.1 Research Framework

4.4 Methodology analysis

4.4.1 Magnitude scale conversion

Different kind of earthquake magnitude type is a result from change in instrumentation, the data reduction method, and the magnitude formula, the station distribution, etc. It need conversion phase, where types of magnitudes converted into same scale. Every area has different conversion analysis formula, based on the number of earthquake events, and magnitude size range. According to Irsyam et. al. (2010), in the research for revision of seismic hazard map of Indonesia in 2010, they analyzed the earthquake catalogue from some sources to be converted using regression analysis. Based on that survey, the result is shown in Table 4.4.

Table 4.4 Conversion correlation

Conversion Correlation `	Numbers of Data (Events)	Data Range	Suitability
$M_w = 0.143\ M_s^2 - 1.05\ M_s + 7.285$	3.173	$4.5 \leq M_s \leq 8.6$	93.9%
$M_w = 0.114\ m_b^2 - 0.556\ m_b + 5.560$	978	$4.9 \leq m_b \leq 8.2$	72.0%
$M_w = 0.787\ M_E + 1.537$	154	$5.2 \leq M_E \leq 7.3$	71.2%
$m_b = 0.125\ M_L^2 - 0.389x + 3.513$	722	$3.0 \leq M_L \leq 6.2$	56.1%
$M_L = 0.717\ M_D + 1.003$	384	$3.0 \leq M_D \leq 5.8$	29.1%

Source : Summary of revision team of seismic hazard map Indonesia, 2010

According to Kulhanek (2005), for global/regional catalogues, surface-wave magnitude, M_s is preferred to use. For data which also include intermediate and deep-focus events, body-wave magnitude, m_b, has to be used. If processing catalogues with large events ($M \geq 7$), moment magnitudes M_w is preferred to use, to solve the saturation effects. Local magnitude M_L will be used in local catalogues.

In this research it is preferred to use moment-magnitude (compared to body-wave and surface-wave magnitude) to indicate the size of large earthquakes.

Usually for studying the space variations of *b,* it is needed to examine the temporal data stability or stationary. Original data are separated into several sub-sets for different time periods and analyzed separately.

4.4.2 b-value calculation

Using eq.8 in section 2.3,

$$b = \frac{\log_{10} e}{\overline{M} - M_c} = \frac{0.434}{\overline{M} - M_c} \qquad \text{...................................(12)}$$

where $\overline{M}$ is the mean of the observed magnitudes and M_c is the magnitude of completeness in the group of events.

For complete data, numbers of data ≥ magnitude completeness data, both spatial and temporal variations in *b* are examined by defining each sliding-space and each time-window. It is implies a varying number of events for each *b* calculation and consequently each *b* is determined with different statistical significance. Selection of a proper window length is depends on data available. There is also an option to use constant-size windows or constant-number of events in the window. This technique implies different window size as the window moves over the grid in which *b*-values are to be determined. This implicated the time scale (for temporal variation examinations) which now becomes non-linear. Constant-number of events technique may have shortcomings generated by time intervals with low level seismicity. If this is the case, long time windows being applied resulting in window lengths much larger than grid-elements and an undesired strong smoothing effect will take place. The final window size (usually determined empirically) is a reasonable compromise between required resolution and smoothing between grid nodes.

Selection of proper magnitude range, for incremental distributions, is arranged by data scarcity. In general, magnitude range should be small to approximate well continuous magnitudes in $N = 10^{a-bM}$, but also each magnitude group should cover large numbers of data. These two opposing requirements should found a proper compromise.

When studying spatial and/or temporal variations in *b*-values, results must be stable, instead of depend on personal choice of input parameters. Tests with different catalogues time spans, window lengths, magnitude completeness, magnitude sampling, etc is needed.

Cumulative distribution provides a better linear fit since numbers are larger and less degraded by statistics of small numbers. It also solved the problem of designing a proper magnitude range.

4.4.3 Fractal dimension calculation

For calculating fractal distribution of seismicity, eq. 2, 3, 4, 5 are used. But first, selecting the window of observed area, and divided into grids with accuracy of 1º and 2º for each latitude and longitude. Fractal correlation dimension (D_2) and the frequency-magnitude relation b-value are first estimated using the epicenters in the grids, grids which each containing < 50 events are discarded. Center of each grid is taken as the plotting point for making contour maps.

Each grid which contains ≥ 50 events is being used. For two epicenters which previously taken as the plotting point each grids, being calculated its distance using eq.5 from sect.2.2.2. Then the correlation dimension, calculated using eq.3. Fractal dimension D_2 simply obtained by plotting the linear function of log-log of $C_q(r)$ and r. Fractal dimension D_2 is the slope of the graph.

5 RESULTS & DISCUSSION

5.1 Magnitude scale conversion

3158 seismic data record of observed area is taken from USGS catalogue from year 1973-June 2011 (Fig 5.1(a)). Numbers of data records satisfy the minimum requirements for good quality earthquakes data. Here, before any scientific analysis is done, assessing the quality, consistency, and homogeneity of the data was done first. Fig. 5.1(a) shows all data records with varies magnitude scale. The first step is to homogenized magnitude scale guided by Table 4.4. For Global catalogue, such as USGS, seismic data converted into M_w. After homogenized, the seismic data record recede until 3144 data left (Fig 5.1). It receded because of some missing value in the process of converting into $M_{W.}$

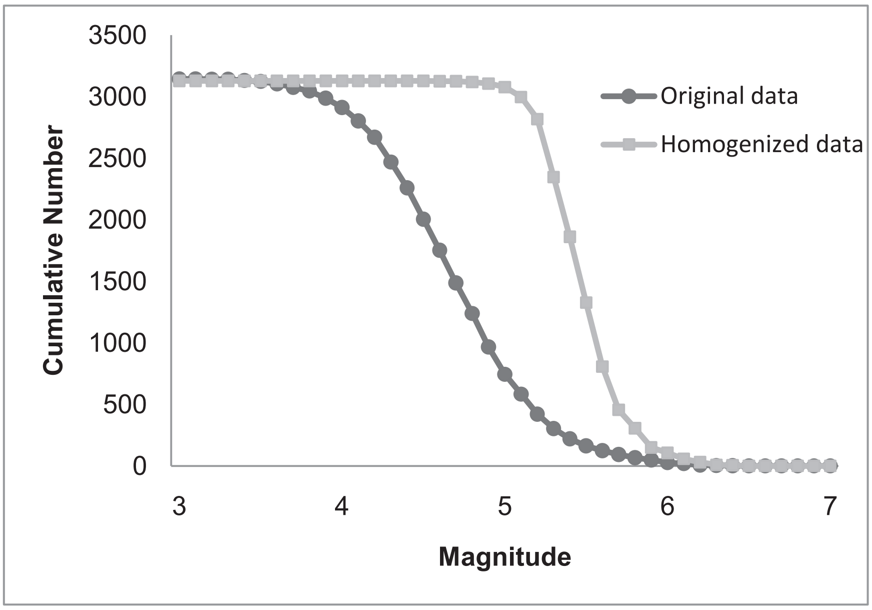


Figure 5.1 Comparison of cumulative numbers of events with varies of magnitude scale/before homogenized and after homogenized

In original data records, magnitude scale range from 3.1 to 6.4 with heterogeneous magnitude type. While the results of magnitude scale homogenization range from 4.5 to 6.4 M_w. After being homogenized, minimum magnitude of completeness M_c is defined, in order to avoid increasing parallel number of earthquake events with time systematically. M_c can be recognized while estimating the frequency-magnitude distribution using varies method.

5.2 b-value calculation & M_C determination

From the data catalogue where numbers of data is 3144, value of $\overline{M}$ is 5,4 M_w. Before the b-value calculated, data are eliminated its depth, where this research only considered shallow and intermediate earthquake, range 0-300 kilometers. Number of data with depth ≤300 kilometers is 3027. Eliminating the earthquake events with depth >300 kilometers purposed that those earthquakes are not considered as threat because its impacts to earth surface.

To calculate b-value, first the level of completeness in the catalogue is evaluated. The level of completeness first assumed that the catalogue simply approximates the frequency-magnitude distribution (eq.(8)). It estimated using two methods; Entire Magnitude Range (EMR), and M_c by b-value Stability (MBS) estimation. The results of frequency-magnitude plot are shown in fig 5.3(a)(b). b-value calculation and estimation are done using Z-map tools software (Wiemer & Malone, 2001).

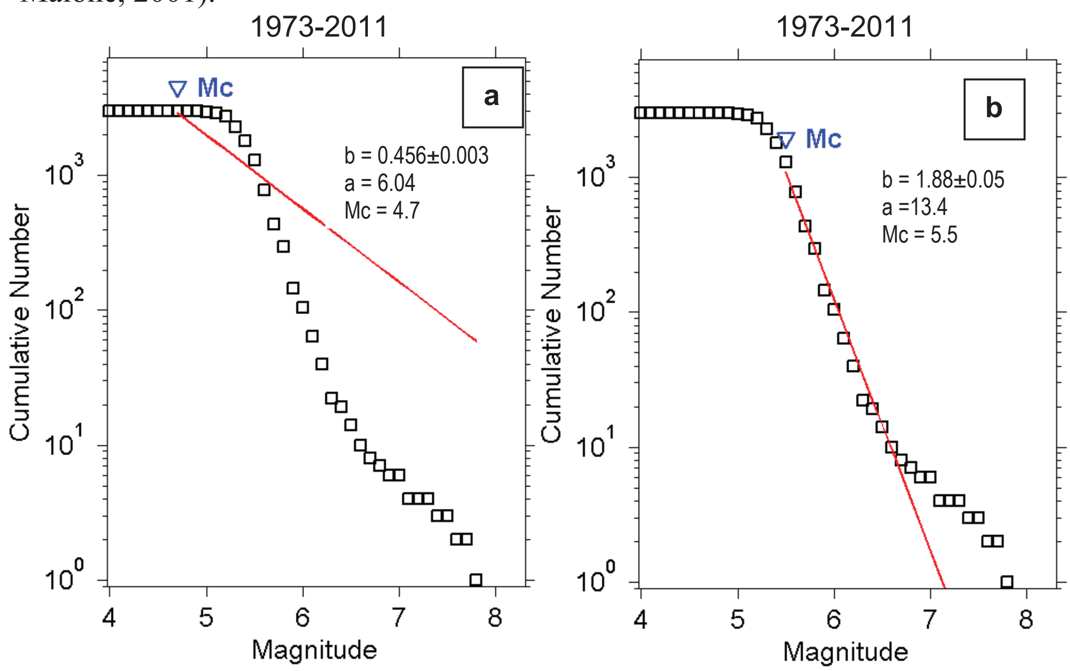


Figure 5.2 b-value estimation & Mc(shown in blue arrow) determination using (a) MBS method (b) EMR method

Using EMR, b-value range is 1.88±0.05. And the magnitude of completeness M_c is 5.5. Using M_c by b-value Stability (MBS) estimation, b-value range is 0.546±0.003. And the magnitude of completeness is 4.7. The higher M_c the more limiting the entire catalogue, so less data analyzed. There are 3023 data left, estimated using M_c 4.7 M_w. For M_c of 5.5 M_w, there are only 1049 left. It seems that M_c 5.5 using EMR method give a better and stable results, also satisfied the Guttenberg-Richter relation. As illustrated in fig. 5.3.

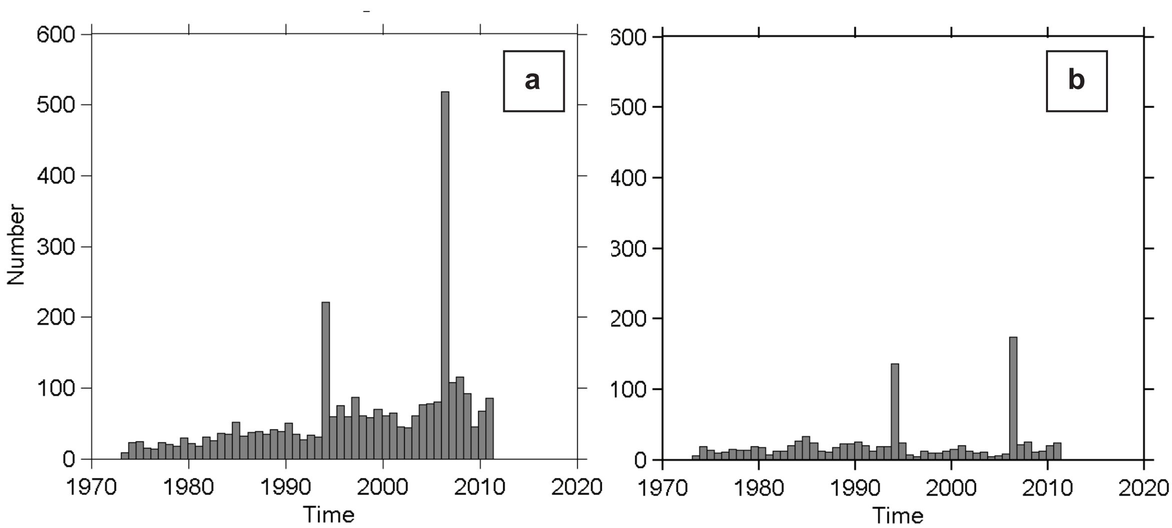


Figure 5.3 Monthly frequency of (a) >4.7 (b) >5.5 Mw.

In general, temporal distribution of the data as shown in fig.5.3(a) & (b) shows that there are two jump in the frequency of seismic activity which means many seismic activity occurring as compared to the other years. One was the earthquake in Banyuwangi, East Java, in year 1994. Other was occurs in year 2006. This is due to the Yogyakarta earthquake, followed by a range of aftershocks, causing great impacts to the whole of the region.

The illustration of monthly frequency in Fig.5.3(b) also shows that data set with >5.5 M_w considered as homogenous and more stable, because the monthly number of them did not increase with time systematically.

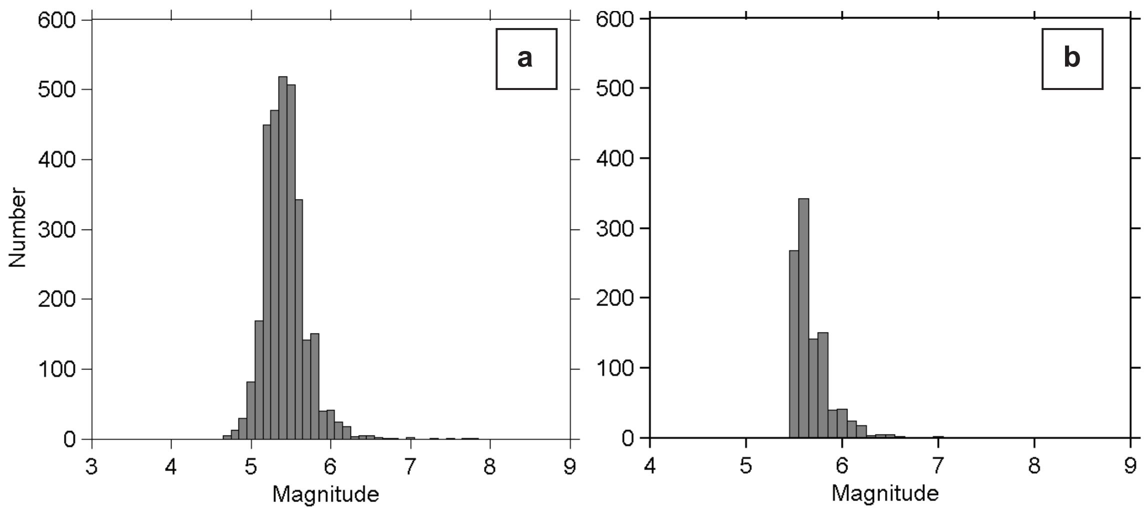


Figure 5.4 Magnitude frequency of (a) >4.7 & (b) >5.5 Mw

Fig.5.4 shows checking results of the distribution as a function of magnitude by plotting the appropriate histogram. From the histogram, it is known that the maximum number is near 5.5 M_w, suggesting that the magnitude of completeness is generally larger than 5.5 M_w, but that it may be near 5.5 M_w in some locations.

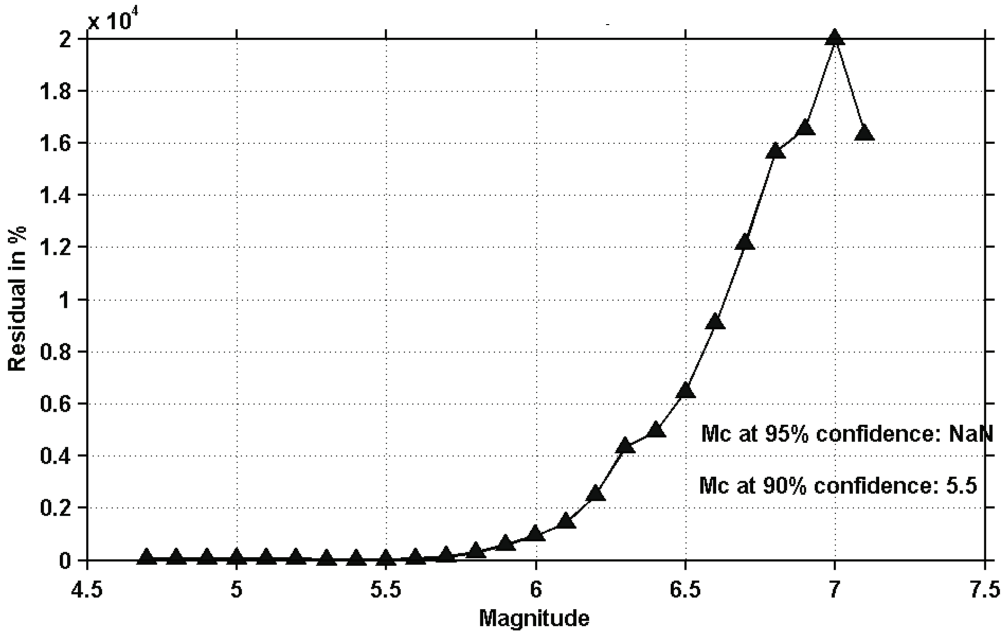


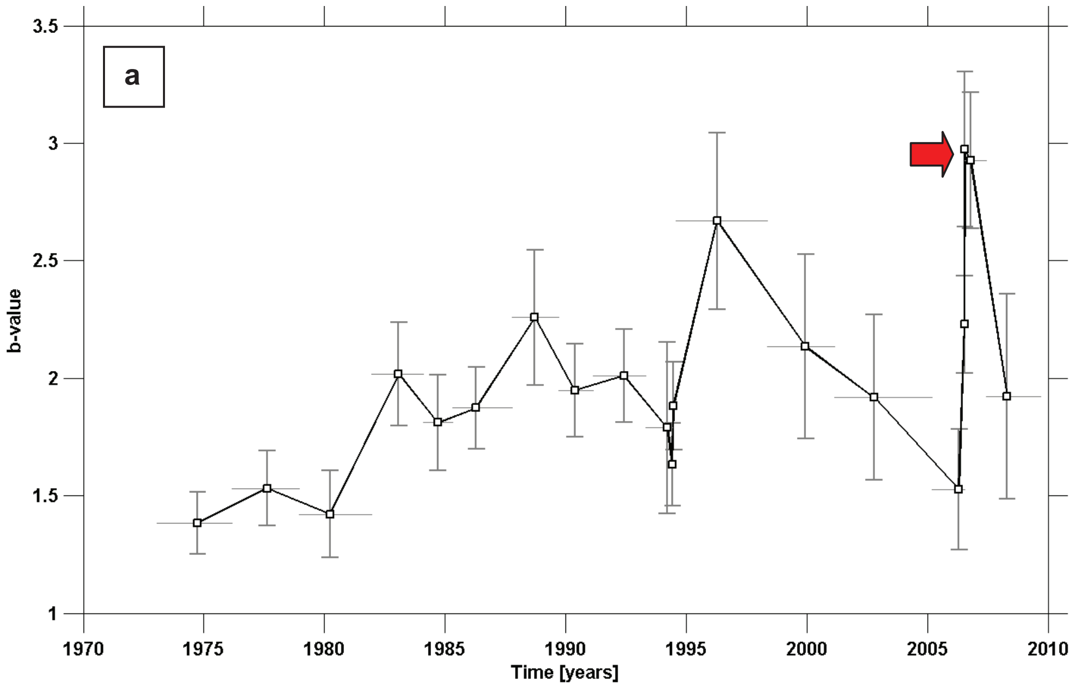
Figure 5.5 Goodness of frequency-magnitude distribution fit to power law

The cumulative Gutenberg-Richter relation (Eq. (7)) is used to evaluate the level of completeness of the catalog, which is considered for further examination using

Eqs. (8) & (10). The cut-off magnitude (M_c) is found to be 5.5 with 90% confidence. From this point, it decided that 5,5 M_w satisfied the Magnitude of completeness (M_c).

5.2.1 Temporal variation b-value

In order to examine the temporal variations in b-value, the sliding time window method applied. As mentioned in section 5.4.2 defining proper sliding time window length will depend on available data. Each time window is composed of 50 or 100 consecutive earthquakes. Data analyzed using M_c 5.5 M_w. Higher M_c, means less data available. From 1049 earthquakes events, there are only 20 time windows for every 50 events, and 10 time windows for every 100 events. See results in Fig. 5.6.



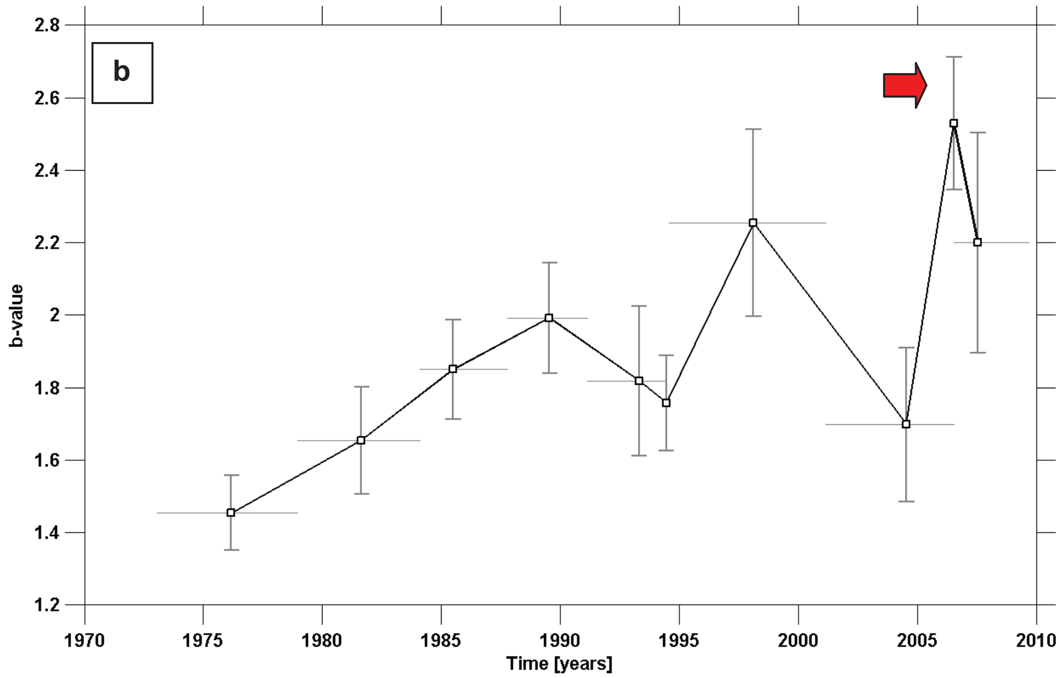


Figure 5.6 b-value using (a.) 50 events each time window (b.) 100 events each time window for *Mc* 5.5 *Mw*

Figure 5.6 (a) and (b) shows temporal variations of b-value for 50 events and 100 events each time window. It shows a significant decrease of b-value from about year 1998 to 2005. The decreasing reach lowest b-value which is about 1,6 for 50 time windows and 1,7 for 100 time windows. Both also indicate that there is a b-value increase after some significant decrease. It is due to the large earthquake event that about to occurs. Sudden increase of b-value in year 2006, indicates the main shock of Yogyakarta earthquake, as shown with red arrows.

Based on the increasing temporal change in cumulative moment, the whole catalogue divided into two-time periods (Fig.5.7). Sub-division 1 is range from year 1994-2001. And Sub-division 2 is range from year 2002-2007.

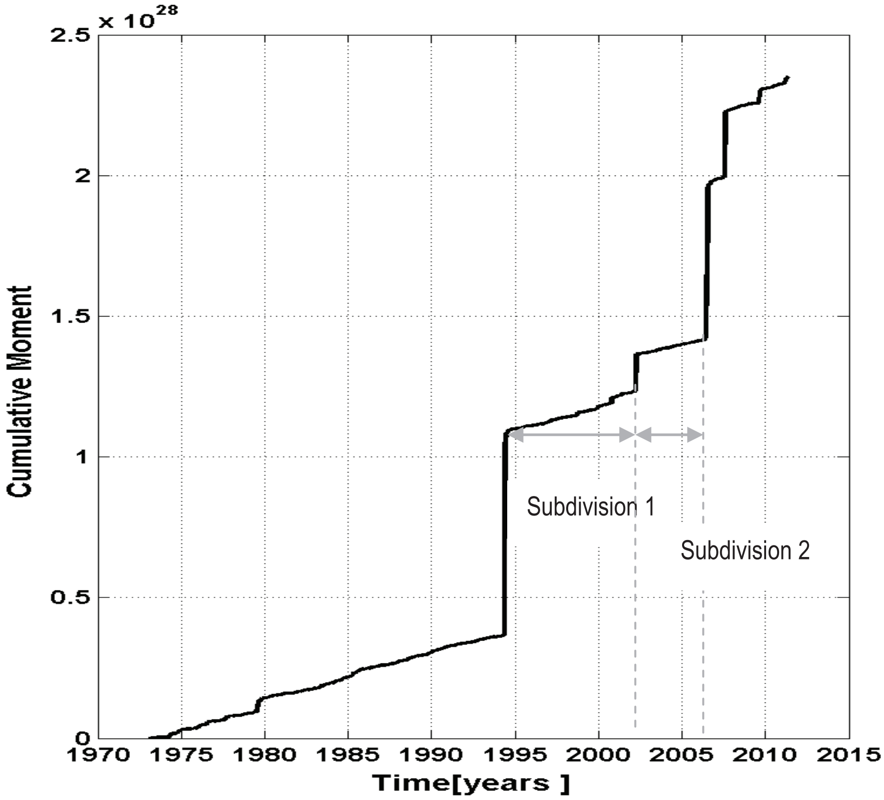


Figure 5.7 Cumulative moment energy release versus time (in years)

These sub-catalogues create based on the sudden increase in seismic moment release in year 2002 and a distinct seismic moment release in 1994 compared to the previous years. The distinct is caused by large earthquake event in 1994, in Banyuwangi with 7,8 M_w. And the sudden increase in 2002 caused by Yogyakarta earthquake in 2006 6,3 M_w, and following aftershocks. Each of sub-division calculated its magnitude of completeness using Entire Magnitude Range (EMR) and M_c by b-value Stability (MBS) estimation (Fig.5.8).

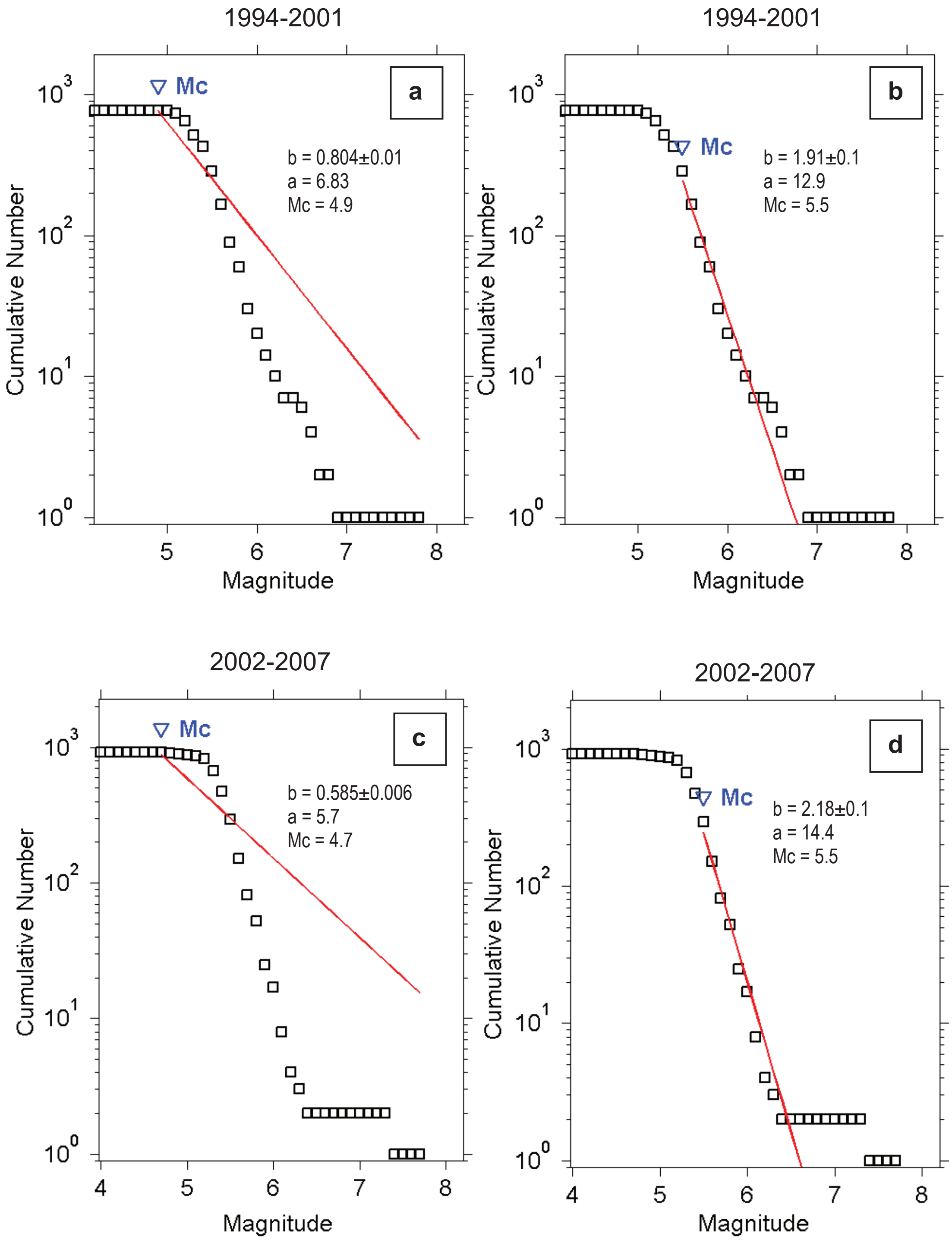


Figure 5.7 b-value estimation using Maximum likelihood method for (a) 1994-2001 catalogue with Mc(shown in blue arrow) determination using MBS estimation method (b) 1994-2001 with Mc determination using EMR estimation method (c)2002-2007 catalogue with Mc determination using MBS estimation method (d) 2002-2007 with Mc determination using EMR estimation method

For catalogue Subdivision 1, year 1994-2001, using EMR, b-value range is 1.91±0.1. Using M_c by b-value Stability (MBS) estimation, b-value range is 0.804±0.01.

For catalogue Subdivision 2, year 2002-2007, using EMR, b-value range is 2,18±0.1. Using M_c by b-value Stability (MBS) estimation, b-value range is 0.585±0.006.

From here, definite irregularity in the behavior of frequency-magnitude distribution in the domain of strong earthquakes is noticeably.

5.2.2 Spatial variation b-value

Determining spatial variation b-value related to earthquake depth, can be seen in Fig.5.8.

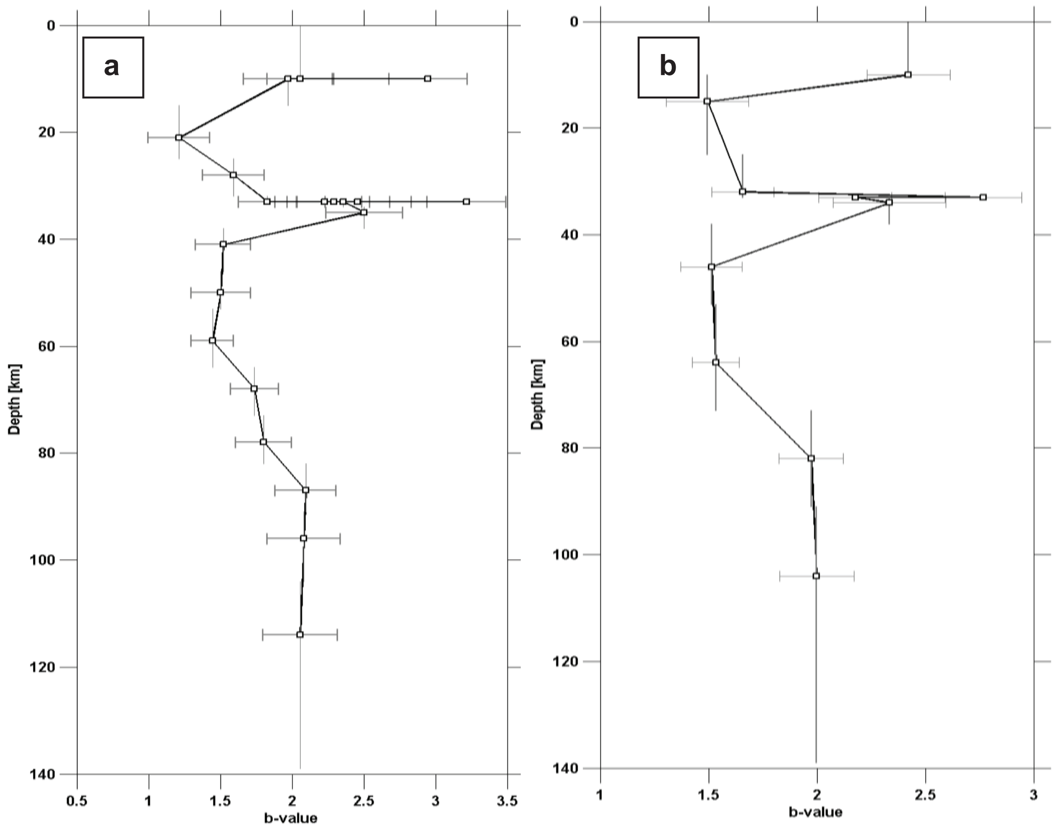


Figure 5.8 b-value variation with depth using (a) 50 window (b) 100 window

As been mentioned in earlier results of Section 5.2, b-value estimate using EMR is 1.88+/-0.05. For mapping visualization of b-value range, the study area range from 5-12 LS and 105-115 BT is gridded with 1º x 1º. This generates 42 grids. The b-value estimated using the epicenters in those grids. Each grid containing more ≥ 50 events is taken its centre as the plotting point for making contour maps. See figure 5.8.

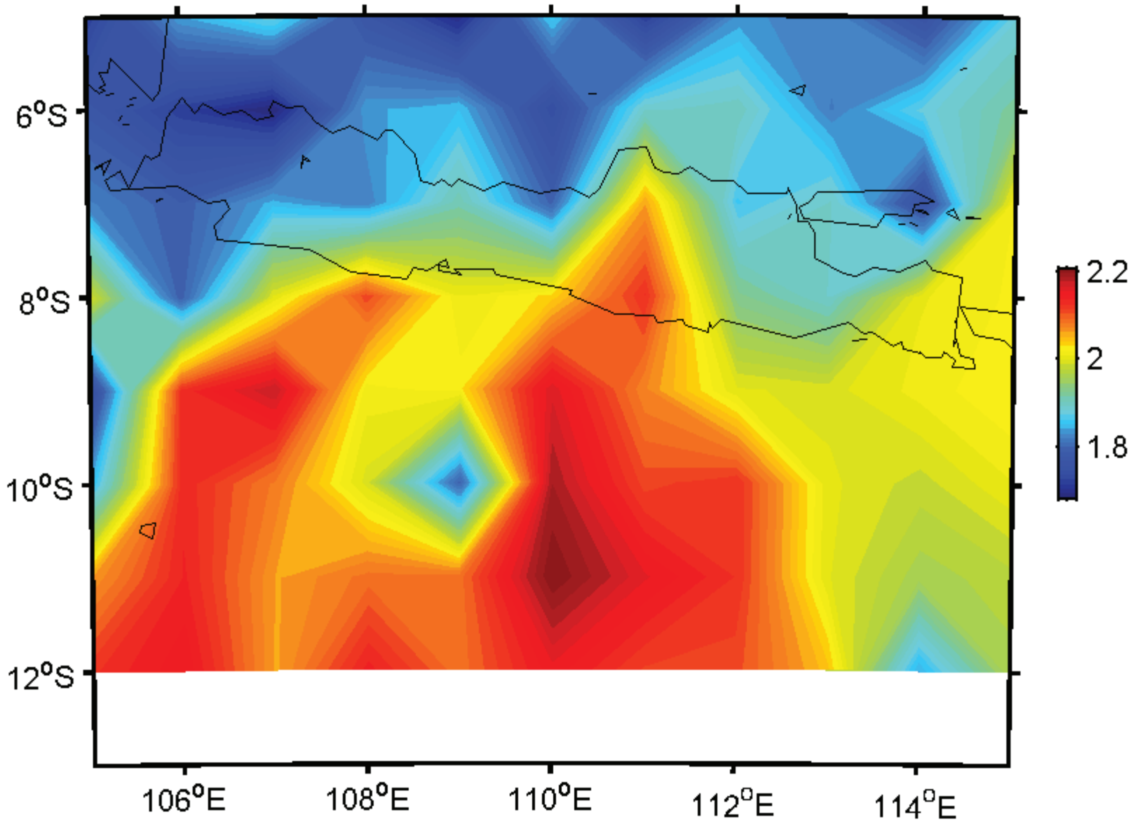


Figure 5.9 b-value map using EMR estimation in 1º x 1º grids

From figure 5.9, it shows that the south of the observed area has higher b-value. While in the smaller observation, Yogyakarta has a b-value range between 1.9 and 2.1.

For mapping visualization of b-value range, the study area range from 5-12 LS and 105-115 BT is gridded with 2º x 2º. This generates 20 grids. The b-value estimated using the epicenters in those grids. Each grid containing more ≥ 50 events is taken its centre as the plotting point for making contour maps. See figure 5.10.

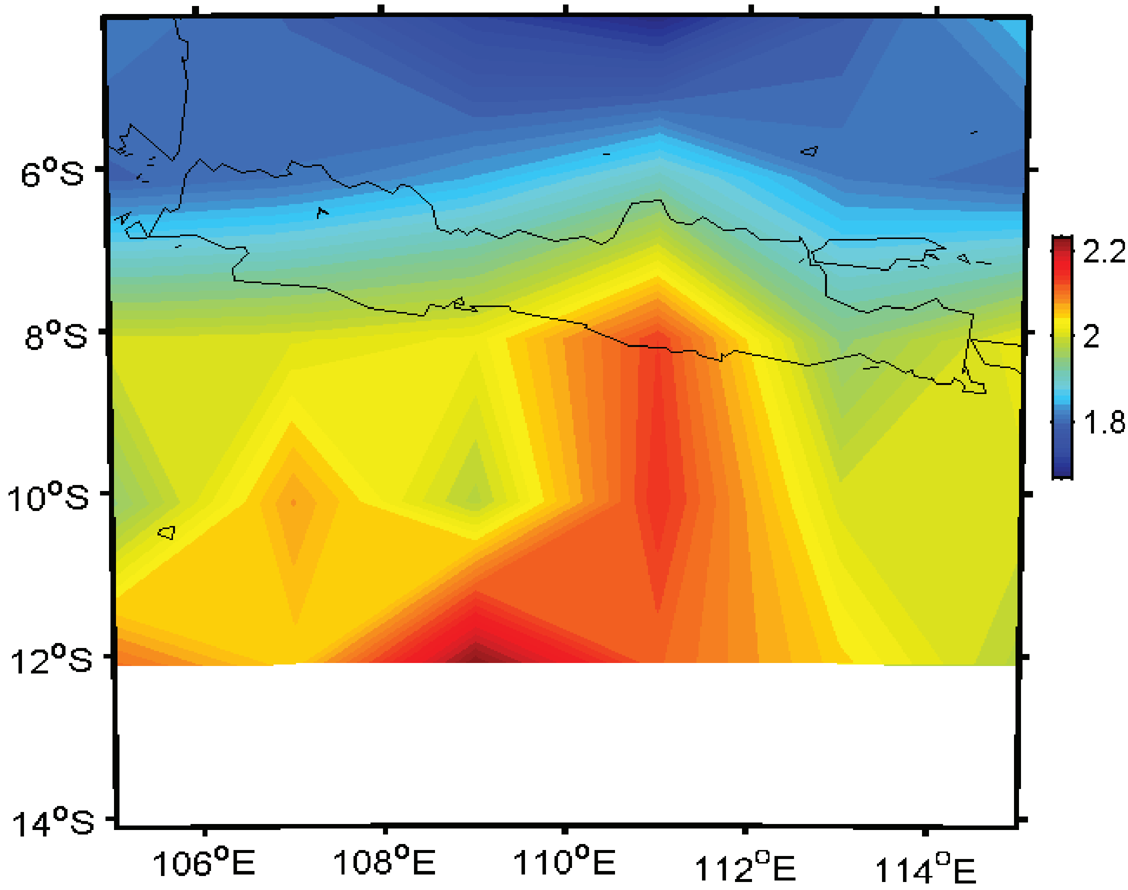


Figure 5.10 b-value map using 2º x 2º grids

From figure 5.10, it shows that the observed area has b-value range from 1.7 to 2.2.

5.3 Fractal dimension calculation

In fractal dimension calculation, both the spatial and temporal variations also analyzed.

5.3.1 Temporal variation fractal dimension

In order to examine the temporal variations in fractal dimension, the sliding time window method applied. As mentioned in section 5.4.2 defining proper sliding time window length will depend on available data. Each time window is composed of 50 or 100 consecutive earthquakes. Data analyzed using M_c 5.5 M_w.

Higher M_c, means less data available. From 1049 earthquakes events, there are only 20 time windows for every 50 events, and 10 time windows for every 100 events. See results in Fig. 5.10.

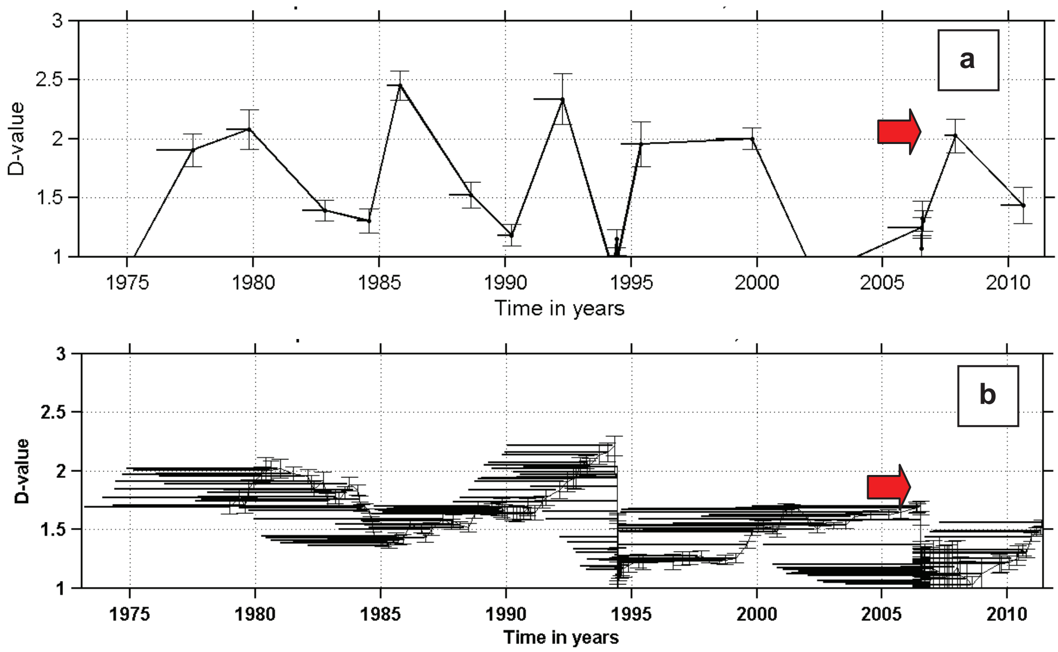


Figure 5.11 D-value temporal variation using (a.) 50 events each time window (b.) 100 events each time window

Fig 5.11 (a) shows a better visualization of D-value changes.

Aki and Tosi (as cited in Roy et. al., 2010) said in their study, fractal correlation dimension value close to 0, means that epicenters clusters into one point; value close to 1, means that line sources are predominant; value close to 2, means that system is being filled up by a plane, and epicenters distributed randomly or homogenously distributed over 2-dimensional embedding space; value close to 3, earthquake fracture are filling up a volume of the crust. A high fractal correlation dimension where epicenter distribution less clustered, indicates that epicenter distribution can be concentrated along the barrier which gives the main shock (Roy, 2006).

Prior to the 2006 Yogyakarta Earthquake, unusual *D* decrease appeared. This change style resembled the cases of other large earthquakes. Therefore, these *D*

changes may be precursors of these large earthquakes. Now, the *D* estimated by the correlation integral method is sensitive to the concentration of hypocenters. If most of earthquakes are in hypocenter clusters, the *D* will decrease effectively.
First, as its cause, the hypocenter cluster is formed by swarm earthquake or aftershock activity which has a small activity area. In this case, it seems that seismic activity becomes active. Second, when a seismic gap is formed by seismic quiescence, such situation may be realized. In this case, it is expected that the earthquakes are isolated and compose clusters, since a seismic gap cuts the spatial connection of hypocenters. In conclusion, both seismic activation and quiescence are effective for *D* decrease.
In this study, it was found that the *D* value began to decrease in 2000, and had been very small for about one year before the main shock occurrence. Such a *D* decrease before the main shock occurrence is a characteristic of some recent large earthquakes. Therefore, the *D* decrease may be an earthquake precursor. The *D* value decrease is yielded by both of seismic activation and quiescence which have often been reported as an earthquake precursor, due to a property of the calculation method. Therefore, the *D* change can be detect, even if the seismic activation and quiescence occur simultaneously in which case the number of earthquakes does not change significantly. On account of this property, using the *D* value is advantageous to detect the precursory change of seismic activity before a large earthquake.

5.3.2 Spatial variation fractal dimension

For mapping visualization of fractal dimension, the cross section A-B is made to see the fractal dimension relation with depths, to illustrate the condition of the area.

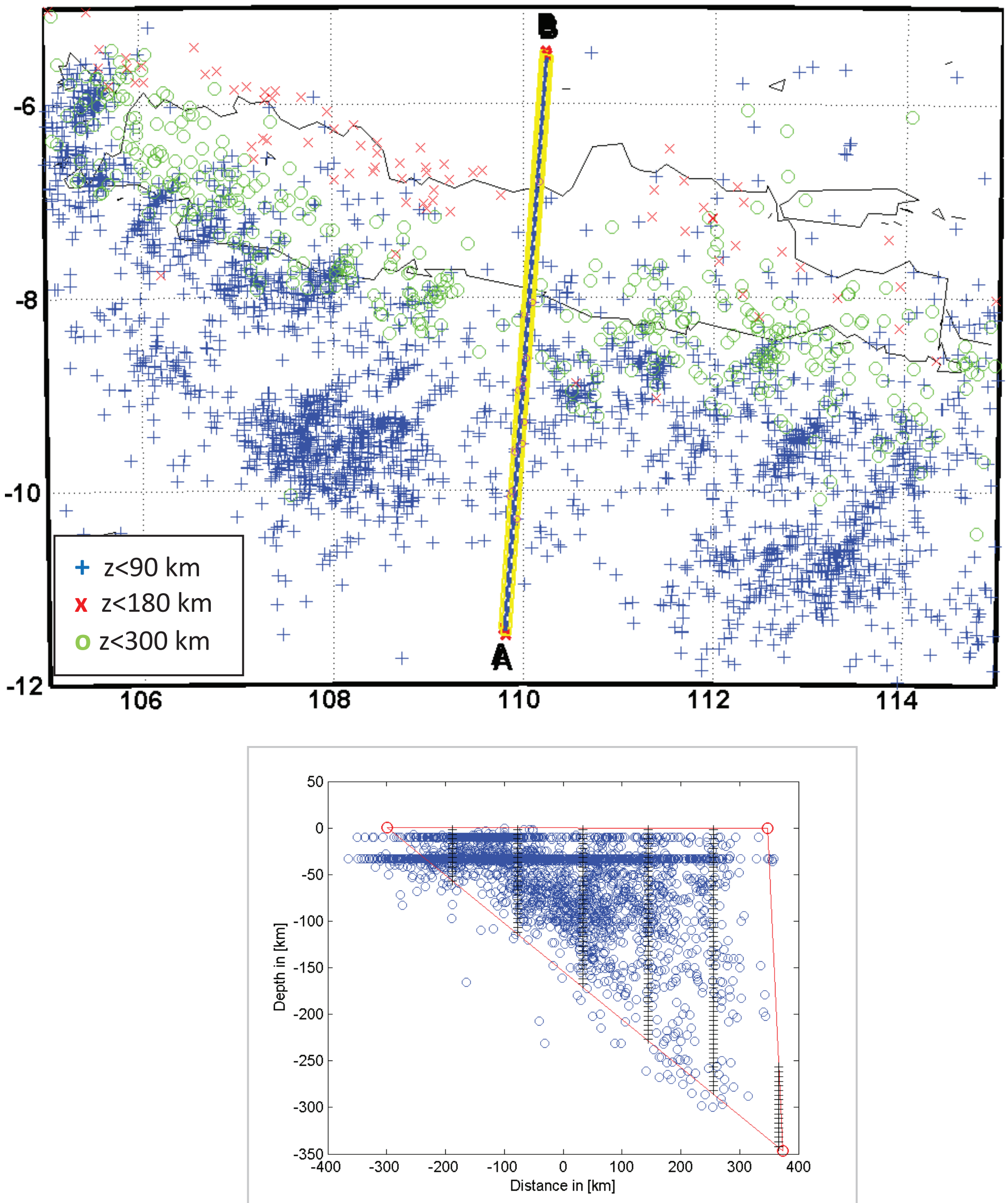


Figure 5.12 Cross section A-B and depth profile

Fig. 5.12 examines the depth profile of cross section A-B with each of 50 earthquake events for every 111 km distance (~1^{o}) and 5 km depth. The red triangle indicate the window observation.

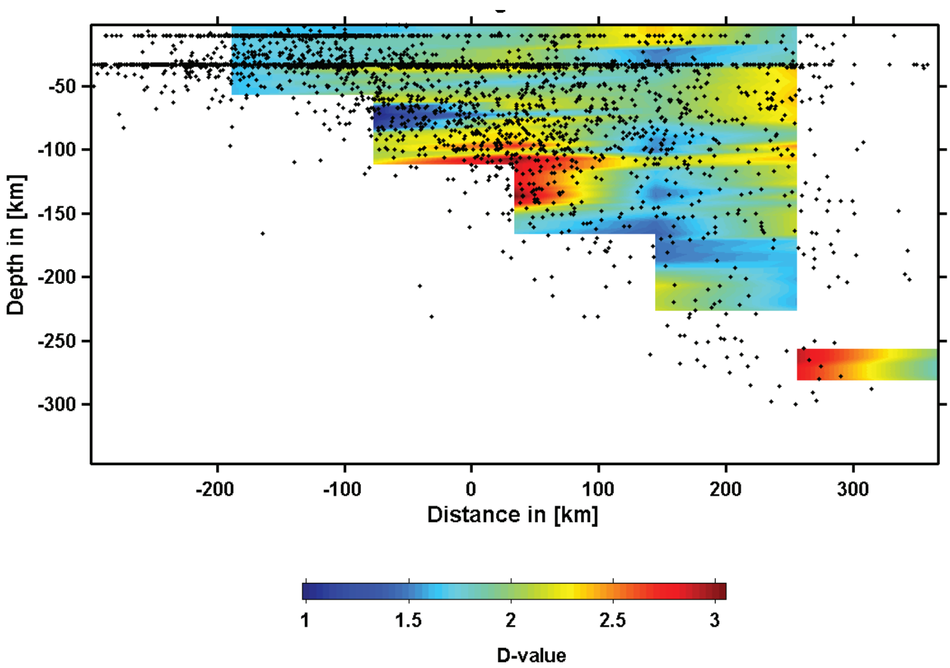


Figure 5.13 Fractal dimension in cross section A-B

Fig.5.13 describes the D-value using weight least square for the cross-section A-B. The spatial variations in fractal dimension D in cross section shows that there is a change in stress of the structure which can be as a mark of a fault movement in the area.

5.4 Fractal dimension & b-value relation

In theory, relation between fractal dimension and b-value is expected to satisfy the Turcotte (1989) formula (eq.12)_$D = 2 * b$. The b-value characterizes the fractal dimension in the domain of energy of earthquakes, whereas D provides a measure of fractal dimension of earthquake location in space. The character of correlation of both, can change. As the result of this research shows that

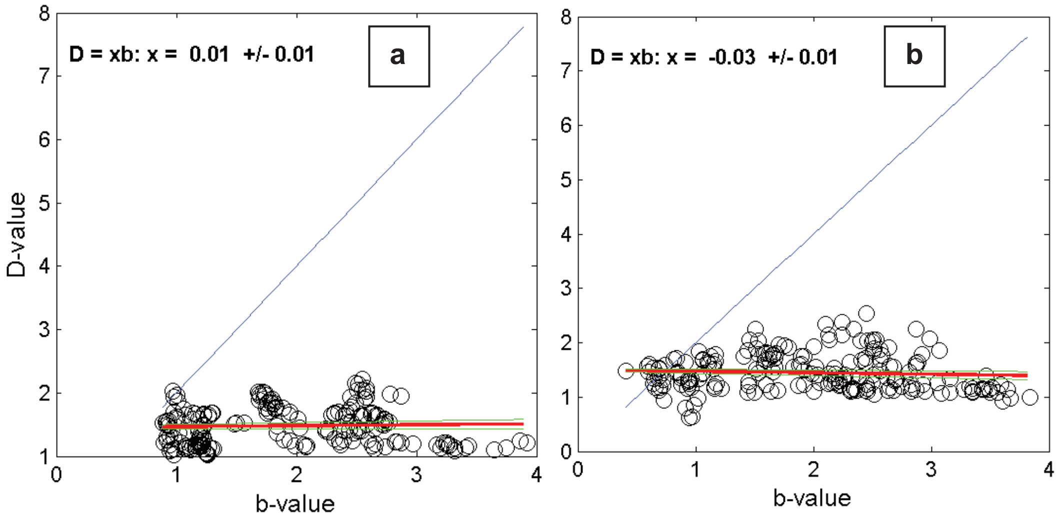


Figure 5.14 Fractal dimension correlation with b-value using (a)50 time windows (b) 100 time window

The fractal dimension correlation with b-value using 50 time windows indicate the positive correlation. While the fractal dimension correlation with b-value using 100 time windows indicates negative correlation.

6 CONCLUSION

The temporal changes in fractal dimension *D* of epicenter distribution before the occurrence of the 2006 Yogyakarta Earthquake was analyzed in this study. The resulting characteristics found by this analysis are as follows:

(1) The value of *D* began to decrease in 2000, and had been very small, for about one year before the main shock occurrence. It is considered that this *D* decrease was a precursor of this earthquake.

(2) The completeness catalogue found to be 5,5 M_w

(3) The value of *D* is decreased by both of seismic activation and quiescence which essentially cancel out each other in the number of earthquakes. It is often difficult to detect either of them simply based on the temporal change in the number of earthquakes. Therefore, the *D* change is advantageous in order to detect an unusual seismic activity preceding a large earthquake.

The spatial variations in fractal dimension D in cross section shows that there is a change in its value which can be an indicator of stress structure change.

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