GEOPHYSICAL CHARACTERISTICS OF THE BOSUMTWI IMPACT CRATER FROM SEISMIC, GRAVITY AND MAGNETIC MEASUREMENTS

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Abstract

Geophysical investigations have been carried out at the Bosumtwi impact crater in Ghana to determine the geophysical characteristics that are related to the impact process. Gravity, magnetic and wide angle seismic reflection and refraction studies have been used to obtain information on the impact-related anomalies. Seismic modelling gave a three layer model of the crater consisting of the water layer with a velocity of 1.45 km/s, post-impact sediments with low velocities of 1.50 - 1.65 km/s and a third layer which is referred to as the crater floor made up of basement rocks. Seismic velocities were found to increase from 2.8 km/s at the interface of post-impact sediments and crater floor to about 5 km/s in 1.6 km depth. The central uplift, which confirms the Bosumtwi crater as a complex impact crater, was found to be 250 m below the water surface. The velocity of 2.8 km/s is interpreted to be due to fallback breccia which completely covered the central uplift. Gravity measurements yielded a maximum negative anomaly of 18 mgal over the crater. This is interpreted to be caused by fractured and brecciated rocks in the rim area and below the crater floor, breccias within the crater, and sedimentary and water infilling of the lake. Magnetic modelling showed that the magnetized bodies are found to be located between 250 and 610 m depths

below the lake. The magnetized bodies have been interpreted to be thin sheets of possibly suevitic impact formations or melt bodies adjacent to the central uplift.

Keywords: Impact crater, Post-impact sediments, Fallback breccias, Gravity anomaly, Magnetized bodies

Introduction

Background Information of the Bosumtwi Crater

The Bosumtwi crater in Ghana is located about 30 km south-east of Kumasi (Fig. 1). The crater is occupied by Lake Bosumtwi and has a rim-to-rim diameter of 10.5 km (Junner, 1937; Jones et al., 1985). It is the youngest large and well preserved impact crater on Earth. The lake itself has a diameter of 8 km and a maximum depth of 75 m (Scholz et al., 2000; Karp et al., 2002). The Bosumtwi crater was formed by a meteorite impact about 1.07 ± 0.05 Ma years ago in lower greenschist facies metasediments (i.e. low grade metamorphosed rocks) of the 2.1 – 2.2 Ga Birimian Supergroup (Storzer and Wagner, 1977; Koeberl et al., 1997a, Wright et al., 1985; Leube et al., 1990).

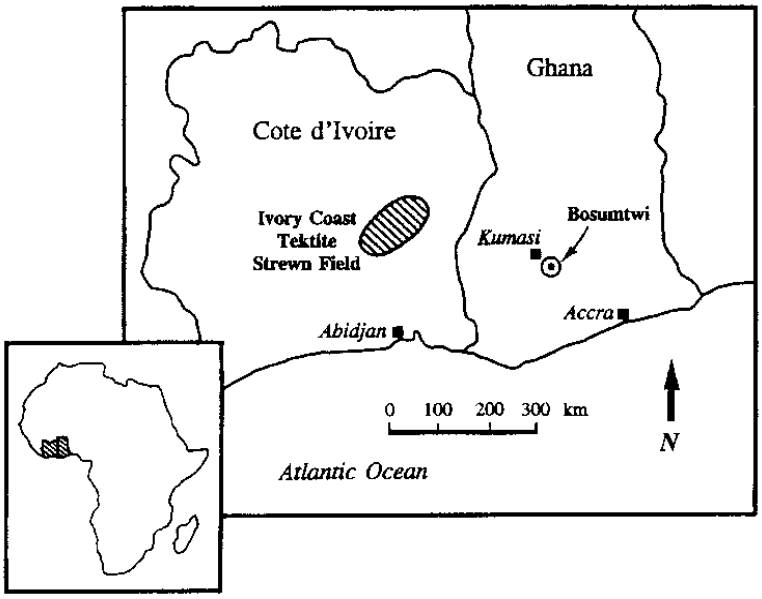


Fig. 1. Overview of the Bosumtwi meteorite impact crater location and relation to the Ivory Coast tektite strewn field (from Koeberl et al., 1998).

Geophysical Characteristics or Anomalies associated with Impact Craters

A major tool in the recognition and confirmation of impact structures on Earth as distinct from volcanic craters is geophysics (Pilkington and Grieve, 1992). The geophysical characteristics of impact craters result from the structural features associated with the formation of the crater. A set of geophysical criteria has therefore been established, from the various geophysical techniques (for example, gravity, magnetic and seismic), that correspond to the geophysical characteristics or signature of impact craters (Pilkington and Grieve, 1992; Grieve and Pilkington, 1996). A number of crater structures were only identified based on the discovery of geophysical anomalies.

Terrestrial impact craters have a set of geophysical characteristics which are largely the result of modification of the physical and lithological properties of rocks by the shock waves, high pressures, high temperatures and the cratering process. The most important are gravity, magnetic and seismic signatures (Pilkington and Grieve, 1992).

Gravity Signature

The most notable gravity anomaly associated with terrestrial impact structures is generally a negative anomaly. When the regional fields are removed, these gravity lows are generally more or less circular and extend to, or slightly beyond, the rim of the structure (Pilkington and Grieve, 1992). The major causes are density contrasts induced by fracturing and brecciation of the target rocks. Minor causes are low-density sedimentary infill of the topographic depression in uneroded structures, and low-density impact-melt sheets in complex structures.

In general, simple craters and small complex craters ($D \leq 10$ km) are characterised by a circular bowl-shaped residual negative anomaly e.g. Deep Bay crater (Fudali, 1979; Pilkington and Grieve, 1992; Grieve and Pilkington, 1996). The amplitude of the maximum negative gravity anomaly associated with impact structures increases with the crater diameter (Dabizha and Fedynsky, 1975; Pilkington and Grieve, 1992). The final character of the gravity anomaly at largely uneroded structures is determined by the diameter D and, to a lesser extent, the pre-impact density distribution of the target rocks. Post-impact processes, such as erosion may cause further changes in anomaly shape and size. For large complex craters, the gravity low may be modified by the presence of a central gravity high which, in a few cases of eroded craters, can be greater than the background field, for example, Vredefort, South Africa (Pilkington and Grieve, 1992).

Magnetic Signature

Magnetic anomalies associated with many impact structures are generally more complex than associated gravity anomalies, and reflect the greater variation possible in the magnetic properties of rocks. The dominant effect in many structures, however, is a magnetic low or subdued zone (Dabizha and Fredynsky, 1975; Clark, 1983), which is commonly manifested as a truncation of the regional magnetic fabric. At larger structures, the magnetic low can be modified by the presence of shorter-wavelength large-amplitude localised anomalies which usually occur at or near the center of the structure. Like the gravity signature, the magnetic signature does not reflect a one-to-one correspondence between the cross-sectional shape of the anomaly and the morphology of the impact structure. Furthermore, the existence of a central gravity high does not imply the presence of a central magnetic anomaly. The nature of the anomaly, whether low or high, however, is somewhat dependent on the size of the structure: magnetic lows occur at small structures, and central high amplitude anomalies at larger structures (about 20 km diameter).

The causes of magnetic lows at impact structures are not as clear (Grieve and Pilkington, 1996). The impact process undoubtedly results in a reduction in the magnetization intensity of the target material. For uneroded craters, post-impact sedimentary infill will tend to be non-magnetic and so contribute to the reduced field intensity. This does not, however, explain the low fields over heavily eroded structures. By analogy with the gravity signature, an important contribution to the magnetic field must come from the parautochthonous target rocks beneath the floor of the structure. Recent studies of drill cores from several Canadian structures indicate that all impact lithologies show a reduction in both induced and remanent magnetization levels, but this is not sufficient to produce all of the observed magnetic lows (Scott et al., 1995). The fractured target rocks also show diminished magnetization levels at depths well below the crater floor, suggesting that the propagating shock wave is likely the cause.

Seismic Signature

Seismic reflection and refraction measurements provide complementary information to potential-field data and geological observations on the characteristics of terrestrial impact structures. Reflection surveys, especially in sedimentary targets, allow for detailed imaging of the crater morphology, and for delineating seismically isotropic zones and incoherent reflections that are characteristic of brecciation and fracturing (Gorter et al., 1989; Grieve and Pilkington, 1996). For example, detailed information on the near surface structure of the Ries crater, where the target is sedimentary rock, has been obtained by seismic reflection measurements (Angenheister and Pohl, 1969). The disturbance of coherent subsurface reflectors is most prominent in the central uplift of complex structures and decreases outward and downward from this zone (Brenan et al., 1975). The depths to horizontal reflectors below the crater floor can be used to determine the amount of stratigraphic uplift as well as to provide estimates of morphological parameters such as the dimensions of the central uplift, annular trough, and faulted blocks at the structural rim of complex structures (Brown, 1973).

Radial variation in impact-induced effects is also apparent on reflection seismic profiles. Where the transition between coherent and incoherent reflectors can be located, this provides an estimate of the dimensions of the so-called transient cavity (Melosh, 1989; Juhlin and Pedersen, 1987). The spatial density and penetration of faults decreases outwards from the annular trough at complex structures (Scott and Hajnal, 1988). Towards the rim of complex structures, the moderately undeformed reflectors within downfaulted blocks allow such displacements to be mapped accurately (Brenan et al., 1975).

Refraction seismic surveys have proved useful for mapping the velocity distribution within terrestrial impact structures – specifically, zones of reduced velocities caused by fracturing and brecciation. At simple structures, velocity reductions of up to 50 percent have been measured within the allochthonous breccia lens and sedimentary infill (Millman et al., 1960; Ackerman et al., 1975). In addition, at the Barringer crater in Arizona, the mapped lower-velocity zone extends out beyond the rim of the structure. At complex structures, the low-velocity zone may extend well below the crater floor e.g., at the Ries crater in Germany (Pohl et al., 1977).

Previous Geophysical Studies at the Bosumtwi Impact Crater

Previous knowledge of the geophysical characteristics, and therefore the structure of the Bosumtwi impact crater, was fairly limited. Because the central part of the crater is buried under water and lake sediments, and because of the lack of drillings, geophysics has to be used to investigate its subsurface structure.

Reports of some general geophysical studies of the area are available (Jones et al., 1981), but no detailed geophysical measurements had been carried out on land and the lake itself. The first modern geophysical work on the lake took place in 1960 when an aeromagnetic survey across the crater area was flown at an altitude of 200 m by Hunting Surveys Ltd. for the Ghana Geological Survey Department as reported by Jones et al. (1981). The occurrence of a central negative magnetic anomaly was detected in this early survey, and was attributed to a breccia lens below the lake sediments. This was followed by the collection of gravity data around the lake as part of a regional survey of south-west Ghana, perfomed by Hastings in 1960 for the Geological Survey Department of Ghana (Jones et al., 1981). Due to the sparseness of data (none on the lake), the gravity data reflect regional trends only. Interpretation of these early measurements (both aeromagnetic and gravity) yielded so far only a hypothetical cross-section of the Bosumtwi impact crater (Jones et al., 1981).

In 1997, a high resolution airborne geophysical survey at an altitude of 70 m was carried out across the Bosumtwi structure by the Geological Survey of Finland in cooperation with the University of Vienna, Austria and the Ghana Geological Survey Department (Ojamo et al., 1997; Pesonen et al., 1998, 1999). The measurements included total magnetic field, electromagnetic field, and gamma radiation. The results of this survey, which provide a more detailed image of the crater structure than the previous measurements, indicate the presence of magnetised bodies, probably suevitic impact breccias or impact melts in the central area of the lake (Pesonen et al., 1998; Plado et al., 2000). An even detailed picture of the nature and form of the magnetised bodies would be obtained if the measurements are carried out on the lake itself, which is one of the objectives of this research.

Recent Geophysical Studies

Geophysical studies have been carried out on both land and lake to determine the geophysical characteristics of the crater which support a meteorite impact origin. This paper reports on detailed geophysical investigations that have been carried out at the Bosumtwi impact crater to determine the geophysical anomalies that are related to the impact process, and therefore confirm an impact origin of the structure. The possible presence of a central uplift structure is an interesting question, as most craters of this size have either a central peak or some kind of peak-ring structure.

Geology of the Lake Bosumtwi Crater Area

The Bosumtwi impact event occurred about 1 million years ago in a target that consisted of Precambrian crystalline rocks, the 2.1-2.2 Ga metasedimentary rocks in greenschist facies of the Lower Birimian System of phyllites, graywackes, quarzites, sandstones, shale, micaschist, as well as local granites (Jones et al., 1981; Wright et al., 1985; Leube et al., 1990; Hirdes et al., 1996; Reimold et. al., 1998). Upper Birimian metamorphosed basalts and pyroclastic rocks (metavolcanics) occur in the Obuom Range, south-east of the crater. Precambrian Tarkwaian metasedimentary rocks occur to the east and south-east of the crater as well (Moon and Mason, 1967; Wooffield, 1966; Jones et al., 1981).

The regional geology is characterized by northeast-southwest trends with steep dips either to the northwest or southeast. However, variations in this trend, due to folding, have been observed (Reimold et al., 1998). Lithology at and around Lake Bosumtwi is dominated by metagraywackes and metasandstones, but some shale and mica schist are found, especially in the north-eastern and southern rim sectors (Reimold et al., 1997; Reimold et al., 1998). A variety of granitoid intrusions (mainly biotite or amphibole granites) have been mapped by Junner (1937) and Moon and Mason (1967). Small granite intrusions, probably connected with the Kumasi granite, crop out around the northeast, west, and south sides of the lake, the largest at Pepiakese on the northeast side of the crater (Jones et al., 1981). In addition, numerous, but generally less than 1-m-wide, dikes of biotite granitoid at many basement exposures in the crater rim have been observed. The overall granitoid component in the region is estimated at about 2 percent (Reimold et al., 1998).

Recent rock formations include the Bosumtwi lake beds, as well as soils and breccias associated with the formation of the crater (Junner, 1937; Kolbe et al., 1967; Woodfield, 1966; Moon and Mason, 1967; Jones et al., 1981; Koeberl et al., 1997b, and Reimold et al., 1998).

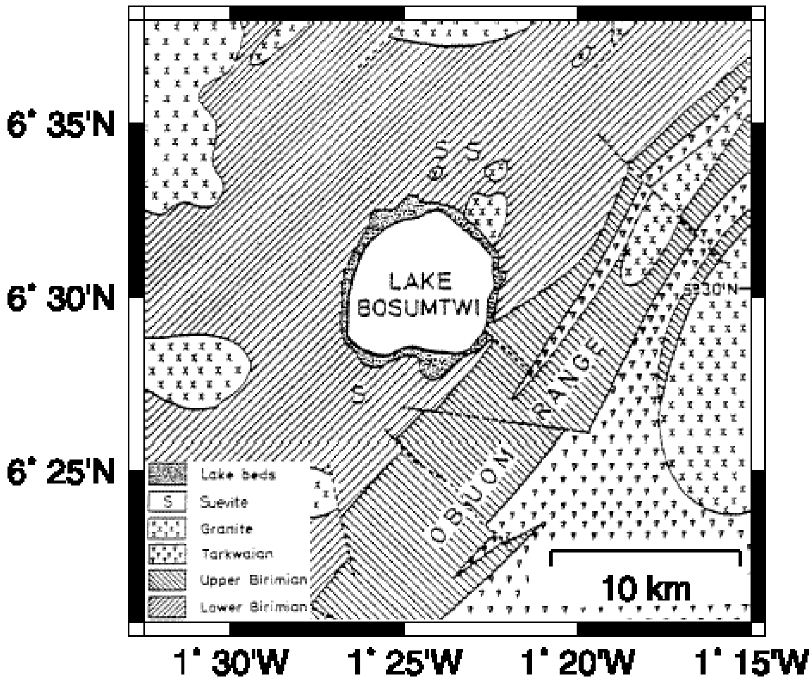


Fig. 2. Schematic geological map of the Bosumtwi crater area (after Jones et al., 1981).

Materials And Methods

Seismic Data Acquisition on Land and the Lake

Wide angle seismic reflection data were acquired on the lake by deploying ocean bottom hydrophones (OBH) at the bottom of the lake to record the seismic signals. The seismic source was a single 0.85l air gun. Shots were fired at intervals of 10 s equivalent to an average shot distance of 25 m. The air gun was towed at a depth of 2 m below the water and behind the motor-driven research platform. Navigation was provided by a differential GPS Zeiss GePos RM 24. In all, the wide angle seismic data was collected along a 6.5 km northwest-southeast profile across the lake (Fig. 3).

Later, additional wide angle seismic refraction experiments were conducted on profiles across the centre of the lake. Ocean bottom hydrophones (OBHs) were deployed at the bottom of the lake, and PDAS seismometers stationed on land, were used to record the seismic signals. Ocean bottom hydrophones labelled OBH 4 and OBH 5 were deployed in the lake while the land seismometers, namely PDAS 1, PDAS 2, PDAS 3 and PDAS 4 were stationed at the following places (see Fig.3):

- PDAS 1: at Wawasi in the far south
- PDAS 2: in an uncompleted building of Dompa Junior Secondary School
- PDAS 3: in a building close to the lake (chalet)
- PDAS 4: at a quarry site outside the crater in the north

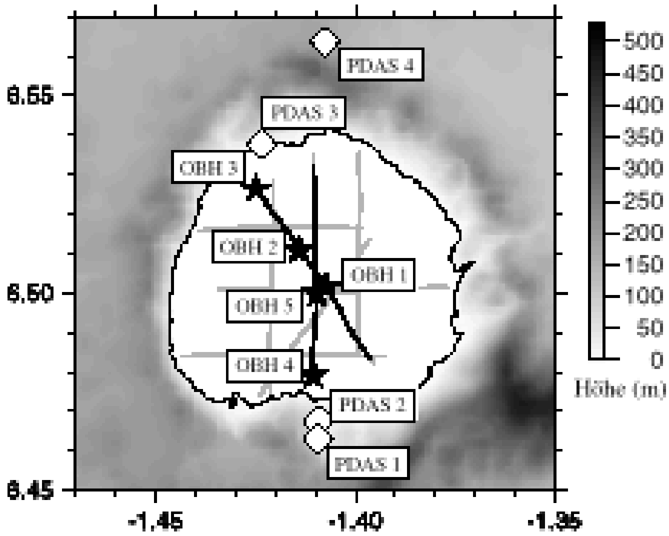


Fig. 3. Location map of the wide angle seismic profiles (black solid lines) and recording stations, OBH (stars) and PDAS (diamonds).

Both the OBHs and PDAS seismometers recorded the seismic signals on a 11-km long North- South profile extending from Wawasi (PDAS 1) through Dompa (PDAS 2) in the south to the quarry site in the north outside the crater (PDAS 4).

Seismic Data Processing and Analysis

All the data (both OBHs and PDAS) were processed using some of the standard routines to improve the signal quality. These include frequency analysis, predictive deconvolution to suppress water bubbles in the case of OBHs data, bandpass filtering, and automatic gain control (AGC), to improve signal-to-noise ratio and to identify the different arrivals.

Gravity Measurements

Gravity Measurements on Land

The gravity measurements were carried out on land including the road linking the 24 villages close to the lake shore, to determine the impact–related crater structure. Elevation measurements were carried out with an accuracy of 2 – 3 cm using differential Global Positioning System equipment (GPS).

For the gravity measurements, a LaCoste-Romberg gravimeter G 256 was used. The instrument has an accuracy of 0.01 mgal. In general, the distance between the gravity points was between 500 – 1000 m. In all, 160 gravity stations were measured. Fig. 4a shows the gravity stations (which are also GPS points) including those of the pillars displayed in the Ghana TM (Transverse Mercator) coordinate system. This was achieved by determining the geographic coordinates (longitude and latitude) of the pillars, and using an affine polynomial transformation to convert these coordinates into the Ghana TM system. The same transformation was applied to all the GPS (gravity points) that were measured in UTM (Universal Time Mercator) system.

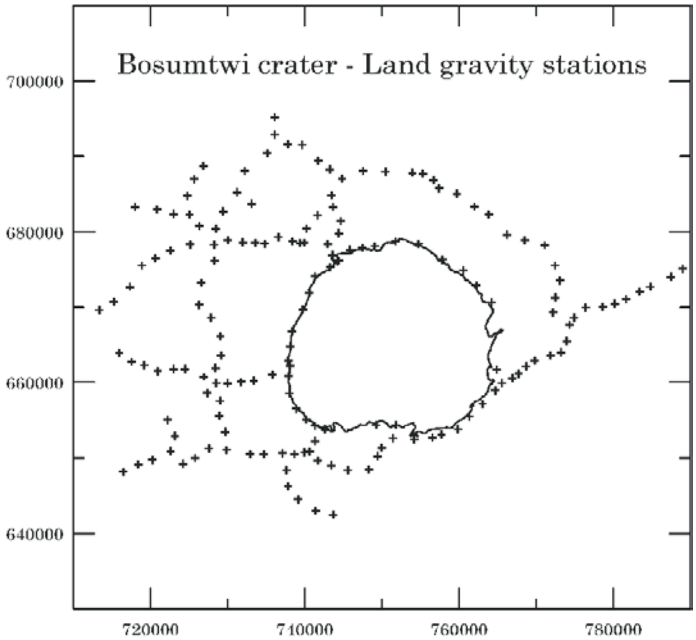


Fig. 4a. Gravity and GPS points around the Bosumtwi crater area.

Gravity Measurements on the Lake

A gravity survey on a grid of 18 profiles, consisting of 11 North-South and 7 East-West profiles, completely covering the lake, was conducted using a LaCoste-Romberg air-sea gravimeter S - 124 belonging to the Geosciences Research Center in Potsdam, Germany. The measurements were carried out using the motor-driven research platform. The distance between the gravity profiles was about 800 m. Navigation was done using a Garmin 235 Echo Sounder/GPS equipment. Fig. 4b shows a plan view of the gravity profiles on the lake.

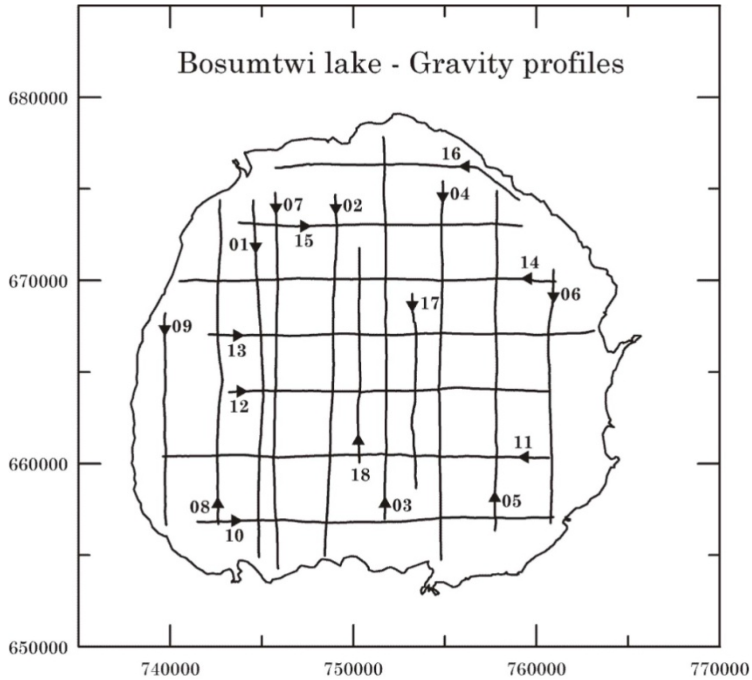


Fig. 4b. Gravity profiles on the lake.

Magnetic Measurements

Magnetic Measurements on the Lake

The magnetic measurements were carried out using two (2) highly sensitive magnetometers GEM GSM-19 TG3. The base station (or reference station) was set up at a quiet location using one of the magnetometers. On the lake, the mobile field magnetometer and sensor were mounted in a rubber boat (unmanned) and towed 50 m behind the motor-driven research platform. Measurements were recorded in automatic mode every 10 s. The lines spacing was 800 m. Navigation was provided by a Garmin 235 Echo Sounder equipped with a GPS. This equipment also provided, at the same time, bathymertic measurements. In total, the magnetic measurements were conducted along 14 north-south and 15 east-west profiles. The data was first processed by removing the diurnal variation of the field. An example of the north-south final dataset is shown in Fig. 8, section 3.2.2.

Results And Discussion

Results of Seismic Measurements

Seismic Signature of the Bosumtwi crater

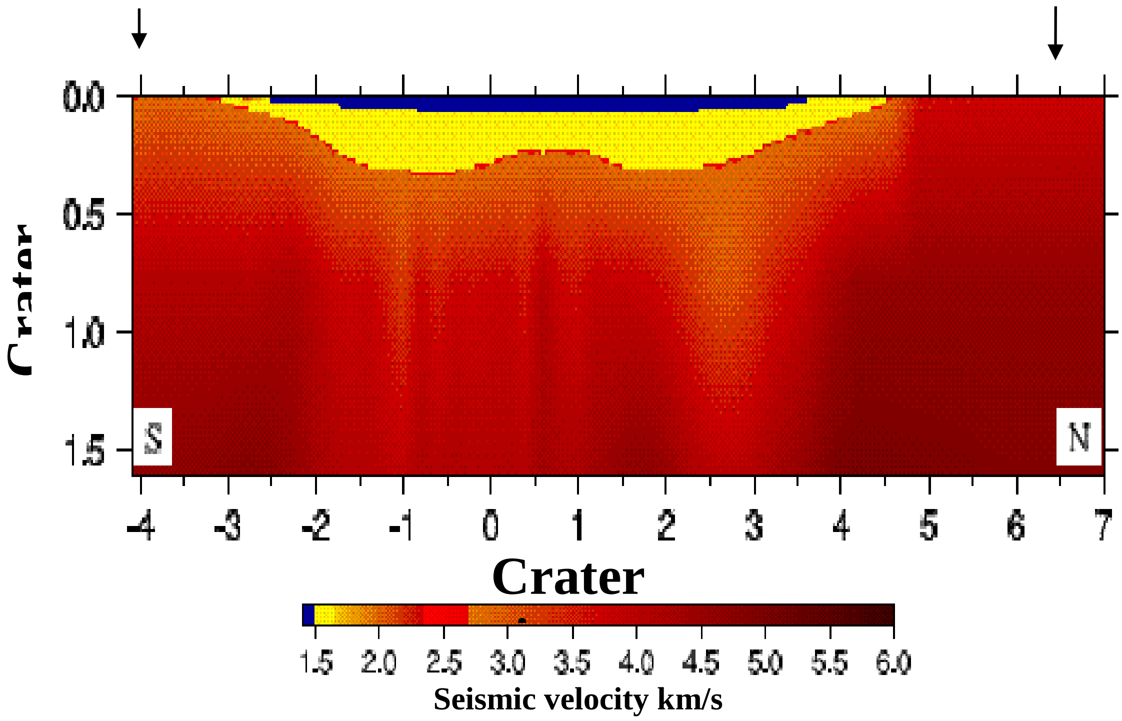


Fig.5. Velocity-depth structure of the the Bosumtwi crater showing a cross-section of the crater along a south-north profile from Wawasi (PDAS 1) to the Quarry Site (PDAS 4), all located outside the crater (vertical exaggeration: 2).

The final result of the ray tracing is a 2-D model (S – N cross section) of the Bosumtwi crater. It is a three layer model with a velocity gradient in the lowermost layer. The seismic measurements using the PDAS seismometers stationed on land (both inside and outside the crater) therefore made it possible to image the deeper structure of the crater down to about 1.6 km depth (Fig. 5).

The model in Fig. 5 had OBH 5 as the reference point. The P-wave velocities were found to be 1.45 km/s in the water and 1.5 – 1.65 km/s in the sediments. The central uplift was found to be 1.9 km wide and had a maximum height of 120 m. The velocity increased in the lowermost layer from 2.8 km/s at the boundary of the post-impact sediments and crater floor to about 5 km/s in 1.6 km depth. The largest velocity value at the maximum depth of 1.6 km is 5.2 km/s and lay between an offset of 4.2 and 5.1 km (Fig. 5). Below the central uplift, the maximum velocity at this depth is 4.7 km/s, but southwards and northwards from it the velocity reduced to 4.2 km/s (i.e. at a distance of -1.0 km from the reference point) and 3.5

km/s (2.5 km from the reference point), respectively. These velocity values are low as compared to values of 5.8 – 6.0 km/s for the undisturbed rocks. The low values are therefore indicative of fracturing of the target rocks.

The seismic velocity in the central area is found to increase with depth, and is interpreted to be due to the presence of uplifted rocks from deeper depths. These were originally found under the transient cavity which formed immediately after the impact, but due to this uplift and the materials sliding down from the steep walls, formed the complex crater after a few seconds.

Around the central uplift of complex craters is a ring-shaped depression (or annular trough), which is covered mostly with rock fragments including fallback breccia and other ejecta material, which were originally found in shallower depths than the uplifted layers in the central uplift (Melosh, 1989). The relative lower velocity at the upper boundary of the crater floor (2.8 km/s) is taken as an indication for such allochthonous breccia layer, which in this case covered not only the ring-shaped depression but also the central uplift. The low velocities to the right and left of the central uplift reached down to greater depths. Whether a particular velocity in this case marked the transition from allochthonous breccia to fractured autochthonous basement rock, could not be explained with the model. However, lateral velocity variations in the depth of 1.5 km were noted, which pointed out that the rock down to at least this depth was affected by the impact and related processes. Investigations of the Ries crater pointed out that the disturbed zone reached down to greater depths (Pohl et al., 1977).

Results of Gravity Measurements

Recent Measurements

For the gravity measurements on land, topographic near-field corrections have been carried out and simple Bouguer anomalies computed. For the lake, the free-air anomalies were computed. Fig. 6 shows a map of the preliminary Bouguer anomalies of the combined results.

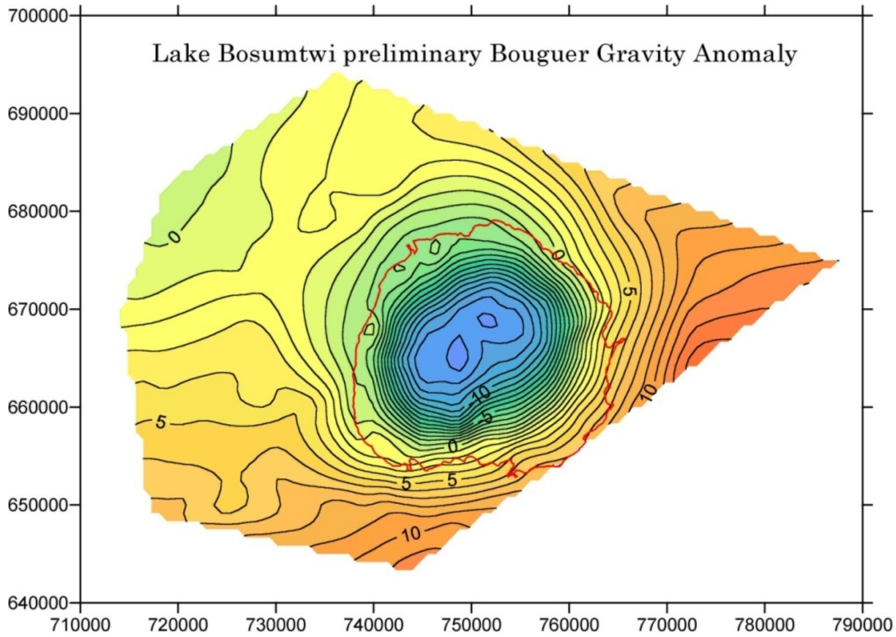


Fig. 6. Preliminary Bouguer anomaly in the Bosumtwi crater area

Gravity Modelling of the Crater

The gravity field of the Bosumtwi area is characterized by a negative Bouguer anomaly with an amplitude of about –18 mgal and a diameter of about 13 km. The steepest gradients are found within the lake area. The central part of the negative anomaly is more flat with a small negative maximum at the center. Contributions to the negative anomaly come from fractured and brecciated rocks in the rim area and below the crater floor, from breccias within the crater, from sedimentary and water infilling of the lake.

Gravity modelling was carried out for a south-north profile across the lake. A linear gravity gradient was removed from the observed data before the modeling. The computations were carried out with the Gravmag programme (British Geological Survey) for 2.5-dimensional geological bodies with half-strike length. Three geological bodies with different densities were assumed: the water in the lake with a density of 1.0 g/cm^3, the underlying sediments with 1.8 g/cm^3 and a breccia layer with 2.0 g/cm^3. The background density value was taken as 2.6 g/cm^3. Information obtained from the seismic results were used to constrain the models. Simple polygons were used as the starting models. Using a forward modelling technique, the gravity response curves of the models were calculated and by changing the shapes of the polygons, the theoretical curves were used to fit the observed data by trial-and-error methods. Fig. 7 is one final model with the observed (solid curve) and calculated (dahed curves) anomalies. A central uplift is clearly shown. It is also observed that the central zone of the lower boundary of layer three at a depth of about 780 m is uplifted.

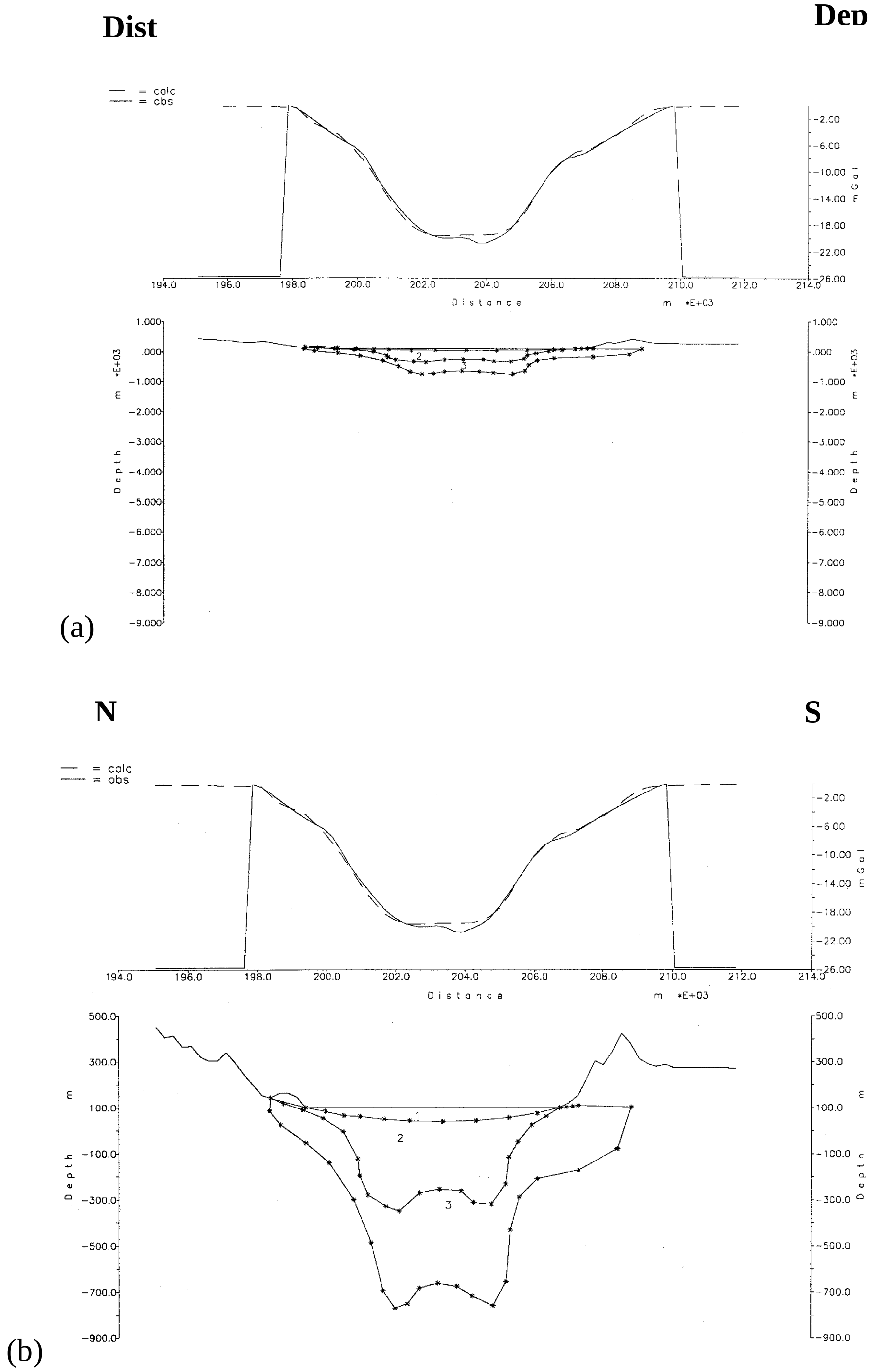


Fig. 7. Observed gravity anomaly (solid curve) and calculated result (dashed curve); (a) without vertical exaggeration; (b) with vertical exaggeration.

Results of Magnetic Measurements

Results of Recent Measurements

The results of the magnetic anomalies along north-south profiles across the lake are shown in figure 8.

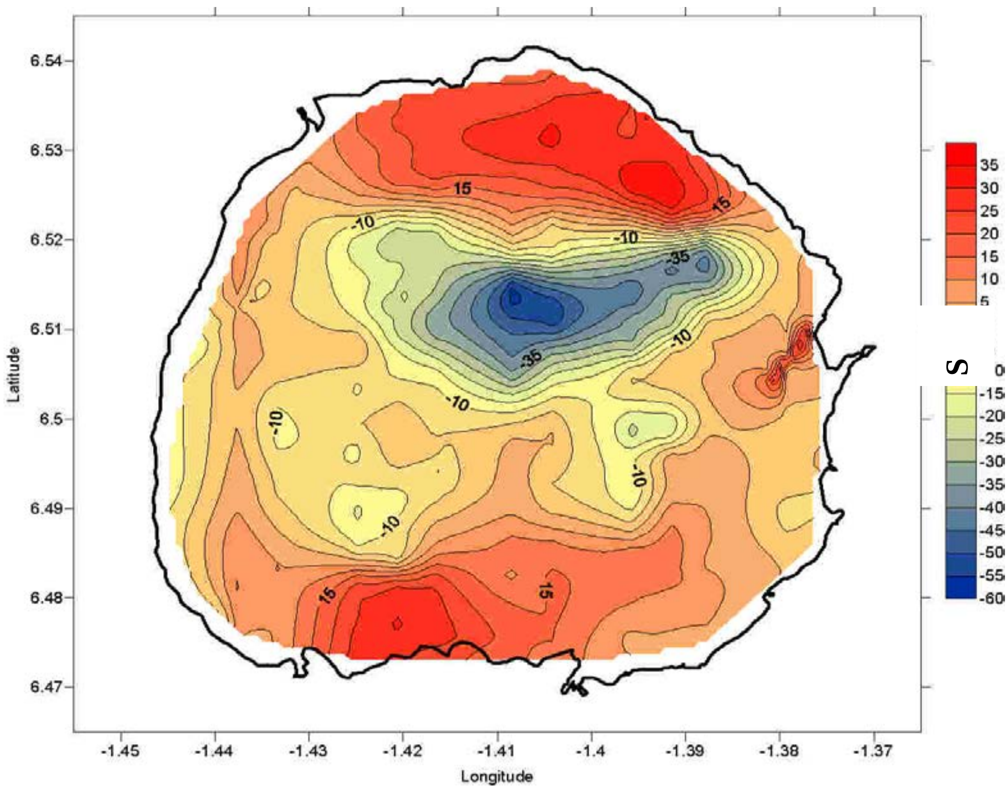


Fig. 8. Magnetic anomalies along north-south profiles across the lake

The main feature of the magnetic field on the lake is a large approximately circular negative anomaly with a minimum value of about 55 nT to the north and a less pronounced positive anomaly to the south. Three to four additional, but weaker negative anomalies are indicated in the southwest and southeast.

Magnetic Signature of the Crater

The shape of the central negative anomaly suggests a normal polarity for the magnetization of the causative bodies (located north of the central uplift) at the latitude of the Bosumtwi lake. Magnetic modelling was done for the main anomaly using the Gravmag software (British Geological Survey) for a 2.5 D model with a half-strike length of 1 km. It is assumed that the anomaly is caused by magnetized bodies formed during the impact process. Evidence of this is found in similar situations observed in many other impact structures, e.g., the Ries crater (Pohl, 1977) and Gosses Bluff (Milton and others, 1972). For the modelling of the data, no regional field had been removed and induced magnetization was assumed. The following parameters of the International Geomagnetic Reference Field were used for the Earth magnetic field, since knowledge of the remanent magnetization for the source bodies is not available (see also Plado et al., 2000):

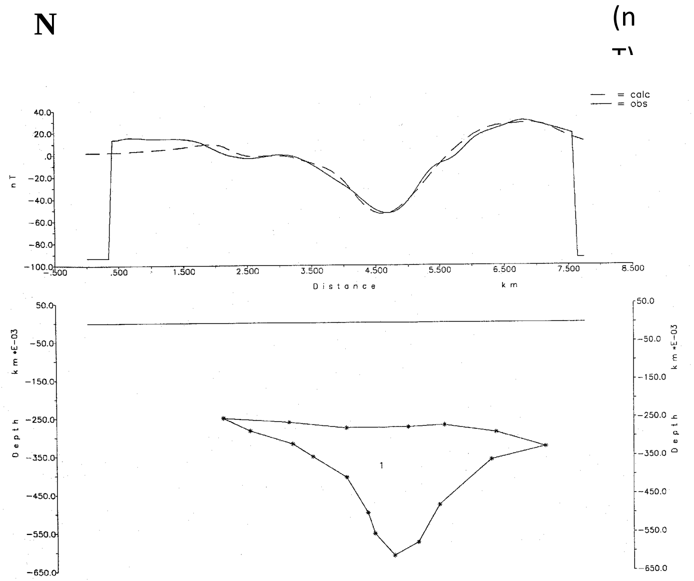
Field intensity = 31860 nT;

Inclination I = -12.5°;

Declination D = 354.4°

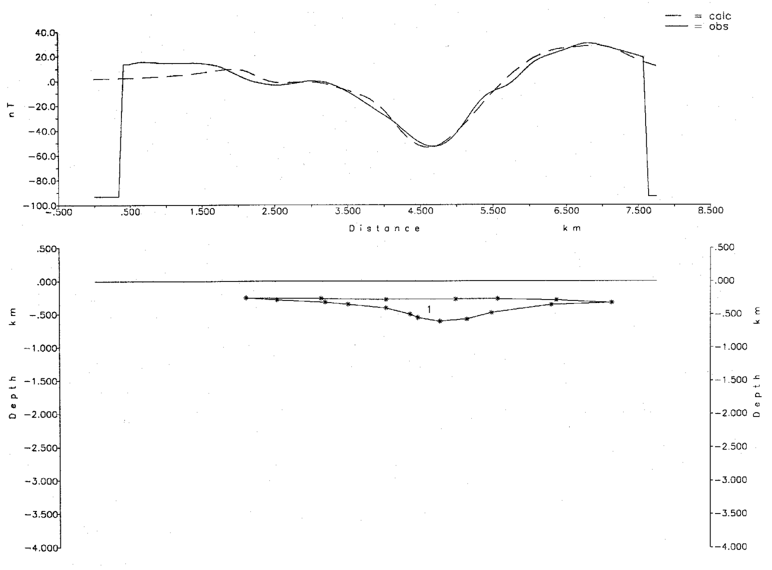
Polygons were used as the starting models. The upper limit of the models was chosen as 250 m in agreement with the top of the central uplift (results of seismic). A forward modelling technique was applied to calculate the theoretical response curves of the models. By changing the shapes of the polygons, the calculated model curve was used to fit the observed data by trial-and-error technique. Some final models obtained for the causative bodies are shown in Fig. 9 (Model 1) and Fig. 10 (model 2). Model 1 has a susceptibility value of 0.03 S.I corresponding to an induced magnetization of 0.075 A/m and the causative body extends down to about 610 m depth, while that of model 2 is 0.05 S.I corresponding to an induced magnetization of 1.25 A/m and the body extends down to about 490 m depth. The relatively high magnetization required for modelling favours interpretation of the magnetized bodies as suevites rather than impact melts.

Model 1: Magnetized Body: 250 – 610 m depth



(a)

Fig. 9a. Observed magnetic anomaly (solid line) and calculated result (dashed curve) for model 1 (with vertical exaggeration).

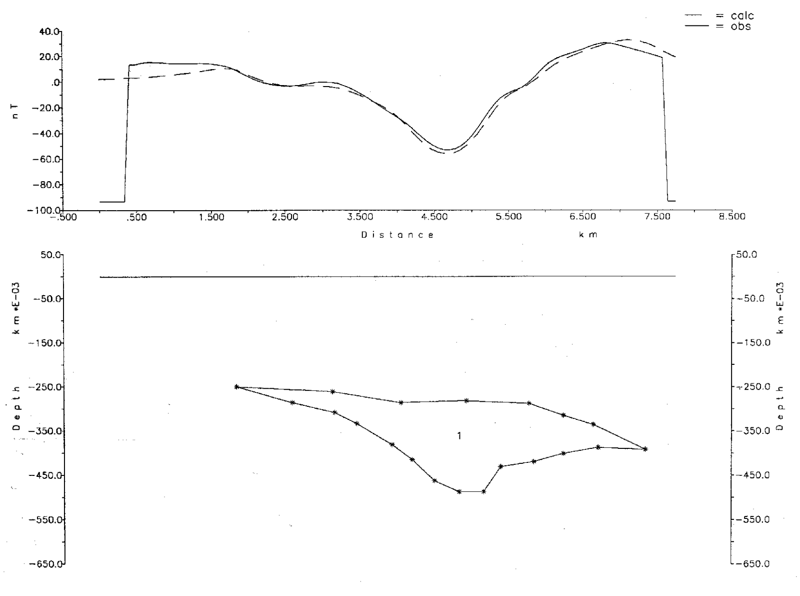


(b)

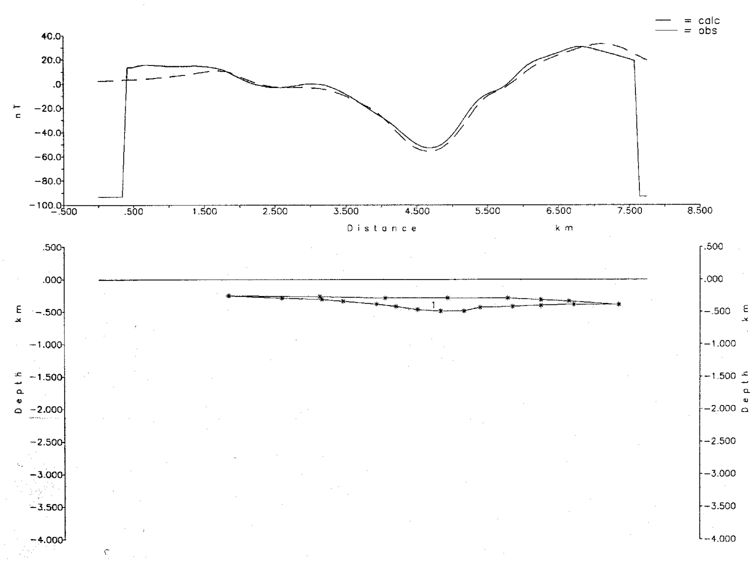
Fig. 9b. Observed magnetic anomaly (solid line) and calculated result (dashed curve) for model 1 (without vertical exaggeraton)

Model 2: Magnetized body: 250 – 400 m depth

N S



(a)



(b)

Fig. 10. Observed magnetic anomaly (solid line) and calculated result for model 2; (a) vertical exaggeration; (b) without vertical exaggeration

The magnetized bodies consist of rather thin sheets of possibly suevitic impact formations or melt bodies adjacent to the central uplift. The weaker anomalies to the west and east of the lake are due to similar bodies. In principle, the results are in agreement with Plado et al. (2000). The central uplift is located at a distance of about 3.97 km along the model profiles Figs. 9 and 10. On the same model profiles (Figs. 9 and 10), the minimum of the magnetic anomaly however lies at about 4.70 km on the northern side of the central uplift. This result is similar to what is observed in the gravity.

Conclusion

The results obtained serve to strengthen the lines of evidence in support of an impact origin for the Bosumtwi crater. The seismic techniques used gave information of the morphology and the nature of the deeper structure of the crater. The crater is found to be a complex impact crater with a central uplift structure. The low seismic velocities of the underlying layers indicate fracturing, shattering and brecciation of the rocks resulting from the impact. Gravity and magnetic results agree with those of the seismics, and indicate the presence of a causative body north of the central uplift. The causative body lies between 250 and 600 m below the water surface and is either melt-rich suevite breccia or impact-melt rock. The results compare favourably well with those of known terrestrial meteorite impact craters.

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