

Petrography and Mineral Chemistry of Ultramafic Xenoliths from Bo Ploi Basalt, Kanchanaburi, Thailand

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ABSTRACT

Ultramafic xenoliths entrained in Cenozoic alkali basalt from Bo Ploi district, Kanchanaburi, were studied for their petrography and mineral chemistry. Then an estimate for geothermobarometry is established. It is found that the entrained ultramafic xenoliths are spinel lherzolite, dunite, and harzburgite. The mineral constituents are essentially olivine, orthopyroxene, clinopyroxene, and spinel. Texturally and chemically these xenoliths are of mantle origin. Their textures are classified as protogranular texture and transitional protogranular to porphyroclastic texture. The coexistence of spinel and plagioclase found in two xenoliths indicates the mantle uplifting to the plagioclase stability field before later ascending magma brought them to the surface. The mineral chemistry of the nodules are characterised by $Fo_{88.3-90.7}$ olivine, $En_{87.1-89}Wo_{1.1-1.2}Fs_{8.8-10.6}$ orthopyroxene, $En_{46.6-48.7}Wo_{45.5-48.6}Fs_{4.2-6.0}$ clinopyroxene, and $Mg\# 0.7-0.8$ spinel. The temperature estimates for these peridotites are in the range of 1012-1189 °C according to equilibrium of enstatite-spinel mineral pairs. In addition, the pressure estimate of 17.6 to 23.8 kbar is obtained, with exceptionally low 11.0 kbar only in one sample. The estimate geothermobarometry reveals the crust-mantle boundary beneath this area is lower than 33 kilometres, where most of the xenoliths are from the depth of 60 to 75 kilometres corresponds to the upper mantle. It is unlikely that these ultramafic xenoliths are cognate material related to the host alkali basalt, hence, the $Mg\#$ of the host alkali basalt, 0.61 - 0.65, is out of range of the $Mg\#$ that is in equilibrium with mantle olivine.

Keywords: mantle xenolith, geothermobarometer, Bo Ploi, Kanchanaburi

1. INTRODUCTION

It is common for Cenozoic continental alkali basalts to entrain mantle-derived nodules that provide direct evidences of the lower crust and mantle compositions, which lead to the better understanding of the dynamics, physical properties, and evolution of the mantle; consequently, geologic processes of the interior earth can be established (Downes *et al.*, 1992, Hsu & Chen, 1998, Jones *et al.*, 2001, Chen *et al.*, 2003)

The Cenozoic basalt in Thailand is believed to be a part of the Late Cenozoic basalt of mainland Southeast Asia (Barr and Macdonald, 1981). The ages of eruption range from 24 Ma to recent. Smith, (1996) proposed that the Late Cenozoic basalt in mainland Southeast Asia had about the same proportion of alkalic and tholeiitic magmas. It is, however, noteworthy that the Cenozoic basalt in Thailand is nearly entirely alkalic affinity while tholeiitic composition has been reported only in a few areas. The wide spread of alkali basalt eruption in Thailand during Late Tertiary to Quaternary is related to tensional regime of continental rift environment that governed by the opening of the Gulf of Thailand and the South China Sea as a consequent of the collision of Indian to Eurasia plates around the end of Mezoic era (Bunopas &

Vella, 1983, Jungyusuk and Khositant, 1992; Smith, 1996). On the other hand, the plumes related hypothesis has been proposed (Barr and James, 1990, Hoke and Campbell, 1995, Smith, 1996). Majority of these Cenozoic alkali basalts have been reported as corundum-bearing alkali basalt and commonly brought with them a variety of crustal and mantle xenoliths with them as in the eruption in Chantaburi-Trat province (Barr&Macdonald, 1981, Jungyusuk and Khositant, 1992, Sutthirat *et al.*, 1999, Sutthirat, *et al.*, 2001), Denchai basalt (Vichit, *et al.*, 1978, Barr&Macdonald, 1979, 1981, Jungyusuk and Khositant, 1992), Ubon Ratchathani - Si Sa Ket basalt field (Barr&Macdonald, 1981, Jungyusuk and Khositant, 1992, Wichianburi basalt (Vichit, *et al.*, 1988, Junyusuk & Sinsakul, 1989, Intasopa, 1993), and Bo Ploi basalt (Barr&Macdonald, 1981, Yaemniyom, 1982, Jungyusuk and Khositant, 1992, Pisutha-Arnold *et al.*, 1998).

The Cenozoic alkali basalt at Bo Ploi, Kanchanaburi, yielded relatively small volume of lava compare to those Cenozoic basalts elsewhere in Thailand. However, it is a well-known locality for mantle xenoliths abundances. Little has been published on the mantle xenoliths from Bo Ploi, Kanchanaburi, and this work report the overviews of the entrained mantle xenoliths in the Bo Ploi basalt in terms of their petrographical feature and mineral chemistry. The results are, therefore, utilized to estimate the paleogeotherm and lithosphere beneath the area can be depicted.

2. Geologic setting

The Bo Ploi basalt is one of the late Cenozoic basalt in Thailand, located in Bo Ploi District, Kanchanaburi Province (Fig. 1). The basalt effuse along the fractures in Silurian-Devonian quartzite (Yaemniyom, 1982; Vichit,

1992) forming several small hills covered an area of approximately 500 m². The basalt has been reported as nepheline-olivine basalt (Bunopas and Bunjitadulaya, 1975), basanitoid (Vichit *et al.*, 1978), and nepheline hawaiite (Barr & Macdonald, 1978; Yaemniyom, 1982). It was given an age of 3.14±0.17 Ma by K-Ar age determination (Barr and Macdonald, 1981) and 4.17±0.11 Ma from ⁴⁰Ar/³⁹Ar age was reported by Sutthirat *et al.*, (1994). Sapphires together with black spinels and pyroxenes have been recovered from weathered basaltic soil, which covered the base of basalt hills as well as from alluvial placer and palaeo-channel deposits.

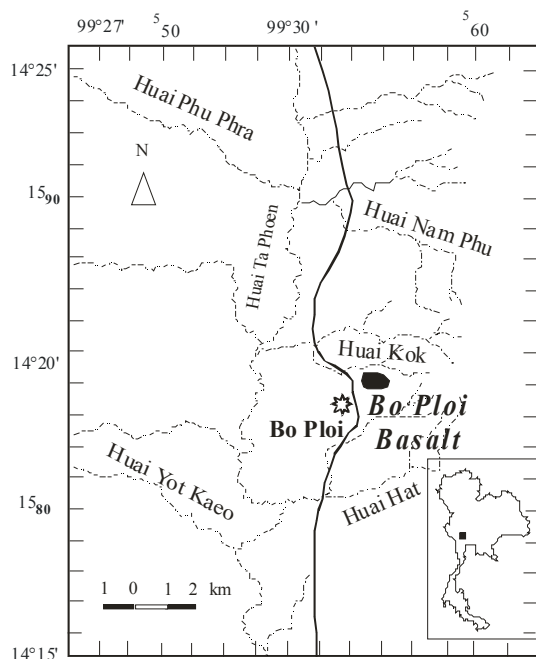


Figure 1 Map showing the location of the Bo Ploi basalt

3. ULTRAMAFIC XENOLITH AND THEIR PETROGRAPHY

Ultramafic xenoliths, crustal xenoliths, and megacrysts are abundantly found in lavas of Bo Ploi basalt. These ultramafic nodules are peridotite. The basalt host rocks are black, dense, fined-grain with obviously porphyritic texture. Phenocrysts and megacrysts or xenocrysts of olivine, black pyroxene and black spinel, glassy sanidine, and, occasionally, magnetite are entrained in the host basalt.

The mantle xenoliths are rounded rectangular to ovoid in shape and range from 1-2 cm across with some larger nodules up to 5 cm across. They are fresh and show a distinct contact with the host basalt. The xenolith primary mineral constituents are olivine, orthopyroxene, clinopyroxene, and spinel. Frequently several grains of these same minerals were carried away in the host basalt lava. Their modal compositions range from harzburgite through lherzolite to dunite (Streckeisen, 1976), though the majority of the xenoliths are spinel lherzolite.

These peridotites exhibit protogranular texture according to the classification of Mercier and Nicolas, (1975) and Pike and Schwarzman, (1977). However, many of xenoliths have deformed texture compare with normal mantle xenoliths; therefore, they are subdivided

into three groups; protogranular texture, transitional protogranular to porphyroclastic texture, and modified texture. Some xenoliths are partly rimmed by tiny olivine crystal layers, which were formed when the xenoliths were incorporated in the basalt.

Olivine, 1-3 mm across, exhibits 'kink band', which is a typical texture observed in mantle olivine. Orthopyroxene has two distinct grain size generations; coarsed-grain up to 5 mm and fined-grain up to 1 mm in size. Coarsed-grain orthopyroxenes frequently show exsolution texture, where clinopyroxene lamellae exsolved from the host orthopyroxene. Deep green clinopyroxene is smaller size, normally show equant grains upto 1 mm. Yellowish brown to dark brown spinel occur in various form, including, bleb-like between pyroxene grains, 'vermicular spinel' resembles exsolution from enstatite, and 'holly leaf spinel' that suggests the forming after recrystallization. Spinel rimmed by plagioclase is observed from those modified texture xenoliths. It is interpreted as a result from decompression, melting and subsequent recrystallization of the xenoliths.

4. MINERAL CHEMISTRY

Mineral chemistry of major phases in the nodules has been determined by a JEOL-733 Superprobe equipped with Link Systems EDS at Birkbeck College, University of London. An accelerating voltage of 15 kV and a count time of 100s have been used. Each sample was randomly selected for 4-6 grains of constituent minerals then each grain was analyzed traversing, 4-6 spots, from rim to core in order to ascertain whether or not any zoning was present since equilibrium compositions are needed for thermodynamic calculation of pressure and temperature. It is found that there are slight chemical variations within certain minerals among these samples.

Olivine has a restricted composition (Fo 88.3-90.7), which is in the range of mantle olivine (Glücklich and Mercier, 1989, Xu *et al.*, 1998, Chen *et al.*, 2003). MnO contents range from 0.07-0.19 wt%, which corresponds to mantle olivine array (Takahashi, 1986). Olivine compositions are shown in Table 1.

Orthopyroxene, enstatite, has near end member composition as $En_{87.1-89}Wo_{1.1-1.2}Fs_{8.8-10.6}$. Mg# falls in the narrow range of 0.90-0.94, which is slightly higher than the coexisting olivine Mg#. Orthopyroxene from protogranular xenoliths have higher Al₂O₃ (mean=5.4) than those xenoliths of porphyroclastic texture (mean=3.8). Orthopyroxene compositions are illustrated in Table 2.

Clinopyroxene is diopside with $En_{46.6-48.7}Wo_{45.5-48.6}Fs_{4.2-6.0}$, and Mg# of 0.91-0.97. Similar to orthopyroxene, Al₂O₃ contents in protogranular clinopyroxene (mean=6.9) are generally higher than those Al₂O₃ of pyroclastic texture (mean=5.6). Clinopyroxene compositions are shown in Table 3.

Spinel has rather variable Cr#, Cr/(Cr+Al), ranges from 0.09 to 0.47. Although most of spinels have Cr/(Cr+Al) ratio in the range of 0.09-0.17, and they are aluminous spinel. A positive correlation is observed when spinel Cr# is plotted against Fo content of coexisting olivine (Fig.2), supporting the idea that these ultramafic nodules are from upper mantle (Arai, 1987, Downes *et al.*, 1992, Unimo and Yoshizawa, 1996). Spinel compositions are present in Table 4.

Table 1 Electron microprobe data for coexisting olivines from Bo Ploi mantle xenoliths

	BPX01	BPX02	BPX03	BPX04	BPX05	BPX06	BPX07	BPX08	BPX09	BPX10	BPX11	BPX12	BPX13	BPX14	BPX15
SiO ₂	41.69	41.49	40.92	41.06	40.96	40.96	40.86	40.95	40.72	41.00	40.85	41.44	41.36	41.37	41.23
TiO ₂	0.02	0.00	0.00	0.00	0.01	0.01	0.00	0.01	0.00	0.02	0.01	0.00	0.01	0.00	0.00
FeO	8.93	10.26	10.35	10.30	10.36	10.41	10.71	10.47	11.28	10.21	10.81	10.04	10.96	9.86	9.46
MnO	0.14	0.09	0.13	0.09	0.13	0.11	0.15	0.15	0.19	0.14	0.12	0.11	0.15	0.11	0.12
CaO	0.02	0.04	0.09	0.12	0.05	0.07	0.06	0.06	0.05	0.09	0.08	0.11	0.05	0.09	0.03
MgO	49.15	48.21	48.59	48.64	48.69	48.67	48.28	48.28	47.84	48.65	48.27	48.26	47.42	48.77	49.42
Cr ₂ O ₃	0.00	0.00	0.01	0.02	0.01	0.03	0.02	0.02	0.01	0.01	0.03	0.01	0.03	0.03	0.01
Total	99.95	100.09	100.09	100.23	100.21	100.26	100.08	99.94	100.09	100.12	100.17	99.97	99.98	100.23	100.27
Fo	90.75	89.33	89.33	89.38	89.34	89.29	88.94	89.20	88.32	89.47	88.84	89.55	88.52	89.81	90.31

Table 2 Electron microprobe data for coexisting orthopyroxenes from Bo Ploi mantle xenoliths

	BPX01	BPX02	BPX03	BPX04	BPX05	BPX06	BPX07	BPX08	BPX09	BPX10	BPX11	BPX12	BPX13	BPX14	BPX15
SiO ₂	56.88	55.15	54.22	53.39	54.93	54.74	54.72	54.89	54.92	54.00	53.90	54.46	55.70	54.57	54.90
TiO ₂	0.09	0.23	0.14	0.19	0.15	0.17	0.14	0.13	0.19	0.15	0.16	0.15	0.12	0.14	0.11
Al ₂ O ₃	2.06	4.38	5.05	6.28	3.84	3.99	4.41	3.81	3.89	5.31	5.23	5.49	3.71	5.12	3.88
Cr ₂ O ₃	0.62	0.45	0.40	0.51	0.40	0.43	0.36	0.41	0.47	0.43	0.43	0.49	0.47	0.52	0.51
FeO	5.73	6.43	6.41	6.36	6.59	6.56	6.43	6.38	6.45	6.38	6.80	6.30	6.48	6.09	5.99
MnO	0.12	0.17	0.14	0.12	0.13	0.15	0.13	0.12	0.12	0.13	0.10	0.10	0.15	0.13	0.53
CaO	0.66	0.85	0.92	1.38	0.56	0.58	0.60	0.61	0.62	0.99	0.96	0.99	0.59	1.00	0.79
MgO	33.62	31.98	32.26	31.27	33.07	32.90	32.79	33.12	33.00	32.07	31.93	31.57	32.19	32.04	33.18
Na ₂ O	0.30	0.36	0.53	0.58	0.45	0.41	0.42	0.46	0.44	0.55	0.51	0.50	0.59	0.38	0.13
Total	100.08	100.00	100.07	100.08	100.12	99.93	100.00	99.93	100.10	100.01	100.02	100.05	100.00	99.99	100.02
Mg#	0.91	0.90	0.94	0.94	0.94	0.93	0.93	0.94	0.93	0.94	0.93	0.91	0.91	0.92	0.93
En	89.96	88.12	88.15	87.1	88.80	88.72	88.86	89.02	88.88	88.02	87.50	88.01	88.61	88.39	88.70
Wo	1.26	1.68	1.80	2.76	1.08	1.12	1.17	1.18	1.20	1.95	1.89	1.98	1.16	1.98	1.52
Fs	8.78	10.20	10.04	10.13	10.12	10.16	9.97	9.80	9.92	10.03	10.61	10.01	10.23	9.63	9.78

Table 3 Electron microprobe data for coexisting clinopyroxenes from Bo Ploi mantle xenoliths

	BPX01	BPX02	BPX03	BPX04	BPX05	BPX06	BPX07	BPX08	BPX09	BPX10	BPX11	BPX12	BPX13	BPX14	BPX15
SiO ₂	53.56	51.85	51.45	51.83	51.25	51.36	51.42	51.22	51.55	51.47	51.43	51.36	52.85	51.95	51.94
TiO ₂	0.28	0.80	0.51	0.53	0.76	0.75	0.63	0.81	0.76	0.57	0.54	0.73	0.78	0.49	0.52
Al ₂ O ₃	3.66	6.00	7.00	7.00	6.22	6.05	6.35	5.86	5.42	7.26	7.17	6.58	4.34	6.52	5.82
Cr ₂ O ₃	1.83	0.88	0.90	0.87	0.98	0.96	0.78	1.03	1.01	0.90	0.82	0.95	1.11	1.06	1.23
FeO	2.30	2.59	3.08	3.19	2.73	2.69	2.65	2.94	2.99	3.19	3.40	3.53	2.46	3.12	2.50
MnO	0.09	0.06	0.09	0.10	0.04	0.05	0.08	0.09	0.08	0.08	0.08	0.10	0.08	0.07	0.09
CaO	21.09	21.54	19.71	19.74	21.45	21.56	21.65	21.51	21.68	19.15	19.52	20.13	21.65	19.61	21.13
MgO	15.69	14.87	15.55	15.07	15.22	15.20	15.25	15.00	15.17	15.67	15.49	15.25	15.17	15.63	15.49
Na ₂ O	1.66	1.42	1.62	1.65	1.38	1.37	1.34	1.45	1.39	1.76	1.71	1.40	1.53	1.53	1.30
Total	100.16	100.01	99.91	99.98	100.03	99.99	100.15	99.91	100.05	100.05	100.16	100.03	99.97	99.98	100.02
Mg#	0.95	0.93	0.95	0.91	0.97	0.96	0.97	0.97	0.96	0.95	0.95	0.91	0.93	0.92	0.94
En	48.74	46.70	49.37	48.48	47.28	47.15	47.15	46.64	46.71	50.11	49.22	48.02	47.18	49.60	48.21
Wo	47.10	48.63	44.98	45.61	47.89	48.08	48.12	48.08	47.99	44.02	44.58	45.57	48.40	44.72	47.27
Fs	4.16	4.67	5.65	5.91	4.83	4.77	4.73	5.28	5.30	5.87	6.20	6.41	4.42	5.68	4.52

Table 4 Electron microprobe data for coexisting spinels from Bo Ploi mantle xenoliths

	BPX01	BPX02	BPX03	BPX04	BPX05	BPX06	BPX07	BPX08	BPX09	BPX10	BPX11	BPX12	BPX13	BPX14	BPX15
SiO ₂	0.29	0.28	0.18	0.21	0.20	0.14	0.23	0.18	0.15	0.20	0.16	0.21	0.26	0.34	0.18
TiO ₂	0.33	0.10	0.15	0.26	0.11	0.11	0.11	0.11	0.15	0.18	0.16	0.19	0.75	0.22	0.12
Al ₂ O ₃	29.62	56.77	57.55	57.14	55.39	55.80	59.53	55.29	55.10	57.13	57.58	57.52	51.86	55.22	53.66
Cr ₂ O ₃	38.61	11.57	10.19	9.76	12.73	12.44	8.82	12.75	12.80	10.14	9.70	11.26	15.51	12.56	14.93
FeO	14.16	10.66	10.79	11.04	11.01	10.98	9.64	10.79	10.84	11.51	11.59	10.04	14.10	11.00	10.78
MnO	0.19	0.11	0.10	0.12	0.15	0.12	0.17	0.13	0.17	0.15	0.13	0.09	0.13	0.13	0.06
MgO	15.35	20.36	20.97	21.40	20.37	20.45	21.57	20.72	20.80	20.79	20.74	20.67	17.74	20.55	20.29
CaO	0.03	0.02	0.02	0.03	0.02	0.01	0.00	0.03	0.02	0.03	0.02	0.01	0.05	0.00	0.00
Total	98.58	99.87	99.95	99.96	99.98	100.05	100.07	100.00	100.03	100.13	100.08	99.99	100.40	100.02	100.02
Cr#	0.47	0.12	0.11	0.10	0.13	0.13	0.09	0.13	0.13	0.11	0.10	0.12	0.17	0.13	0.16

5. PRESSURE AND TEMPERATURE ESTIMATION

Temperature estimation of the mantle xenoliths from Bo Ploi basalt is calculated based on the equilibrium of coexisted orthopyroxene-clinopyroxene compositions (Brey & Köhler, 1990) and enstatite-spinel (Mercier 1980), where the results are summarised in Table 5. The temperature estimate based on two-pyroxene geothermometers; T_{BKN} , gives 919-1144°C, while slightly lower range, 797-1101°C is obtained from Ca-content in orthopyroxene thermometry. Other thermometry worked on equilibrium of enstatite-spinel mineral pairs gives the higher and narrower temperature range of 1012-1189°C (Mercier 1980). Although, there is a variation of estimated temperatures, however, they are systematic.

Pyroxene-spinel geobarometry of Mercier (1980), which works on the role of chromium on aluminium solubility in pyroxene compare to the coexisting spinel was utilized for calculating. The pressure of 17.6-23.8 kbar was obtained from majority of the xenoliths, which is in a good agreement to the stability field of spinel in peridotite (O'Hara *et al.*, 1971). The sample BPX01, however, gives the lower pressure estimate, 11 kbar, which is a result from the high chromium content in the mineral assemblages.

Among these studied samples, some xenoliths do not contain spinel and were not calculated for their pressures. spinel may not actually occur in the spinel lherzolite stability field, if the small amount of Al_2O_3 in the rock and the temperature is close to the solidus. At this point pyroxenes, themselves, allow all the Al_2O_3 on their sites so that the spinel phase does not develop (Hall, 1996).

Table 5. P-T estimate for Bo Ploi mantle xenoliths

Sample	Rock type	Temperature (°C)			Pressure (kbar)
		a	b	c	
BPX01	lherzolite	954	877	1032	11.0
BPX02	lherzolite	1014	855	1079	20.9
BPX03	dunite	1033	1066	1098	22.0
BPX04	lherzolite	1144	1008	1189	23.8
BPX05	lherzolite	919	885	1005	18.3
BPX06	lherzolite	927	866	1012	17.6
BPX07	lherzolite	934	868	1018	18.2
BPX08	lherzolite	938	843	1019	19.2
BPX09	lherzolite	941	832	1022	17.9
BPX10	lherzolite	1052	1101	1114	22.0
BPX11	dunite	1045	1069	1107	21.6
BPX12	lherzolite	1052	1043	1113	20.4
BPX13	lherzolite	930	797	1015	18.2
BPX14	lherzolite	1054	1082	1112	20.4
BPX15	harzburgite	996	962	1063	18.9

a after Brey & Köhler, 1990 (BKN)

b after Brey & Köhler, 1990 (Ca-op)

c after Mercier, 1980 (en/sp)

6. DISCUSSION

6.1 Origin of the ultramafic xenoliths

Texturally and geochemically of ultramafic xenoliths from Bo Ploi basalt indicate that these nodules are of mantle origin. Evidently, MnO of olivine are well defined in the

upper mantle range. The relationship of spinel Cr#, $Cr/(Cr+Al)$, and olivine Fo content also signify mantle origin as illustrated in Fig.2.

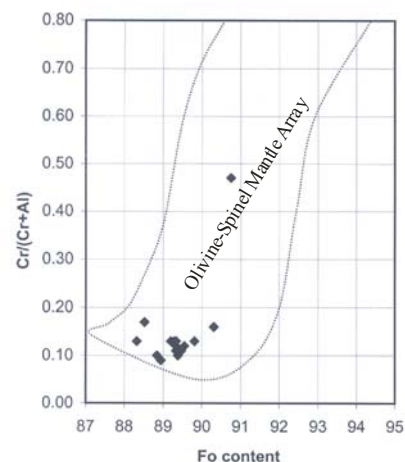


Figure 2 Diagram show spinel Cr# $[Cr/(Cr+Al)]$ plotted against Fo content of coexisting olivine. The dash area, mantle peridotite, is modified after Unimo, *et al.*, 1996

There is a weak relationship between texture and $P-T$ estimation, which may not applicable to all types of xenoliths. Xenoliths of protogranular texture have higher temperature estimates than the other groups. Similar relations were reported from the study of peridotitic xenoliths from young volcanic regions in the Massif Central (Brown *et al.*, 1980) and the xenoliths from Rhenish Massif, Germany (Witt and Seck, 1987). It was suggested by Xu *et al.*, (1998) that this is because the upper mantle may locally constitute layers of undeformed and fertile peridotites, which were overlain by deformed-depleted ones

The narrow variation in temperature estimates (approximately $\pm 100^\circ C$) as well as pressure calculation (approximately ± 3 kbar) among the majority of xenoliths suggest that these peridotites are from similar depth prior the entrainment in the alkali basalt, which correspond to the depth of 60 to 75 kilometres of the upper mantle. The only one sample that gives pressure estimate of 11 kbar corresponds to the depth of about 33 kilometers. The $P-T$ regimes of these peridotites are generally in spinel lherzolite stable field that extends from approximately 8 kbar to 25 kbar (O'Hara *et al.*, 1971, Wilson, 1989).

The coexistence of spinel and plagioclase in sample BPX14 and BPX16, also reported in peridotites from Adamawa, Cameroon (Dautria and Girod, 1987), is interpreted as a result of heating, melting, decompression, and quenching of the xenoliths, while they were incorporating in the host alkali basalt. The xenoliths must be brought to the maximum pressure that was stable for plagioclase to coexist with spinel, which should be at 8-9 kbar. Residing at this pressure would allow plagioclase to crystallize in coexistence with spinel before the later ascending magma brought them to the surface. Alternated explanation for this feature is due to the effect of heat from the host alkali basalt, while the xenoliths were entrained, particularly, for spinels rimmed by plagioclase that located close to the host alkali basalt. The interaction of clinopyroxene and spinel during the melting produces olivine + plagioclase or plagioclase + spinel of a different composition (Pike and Schwarzman, 1977).

The estimated P-T from 11.0 to 23.8 kbar is stable for all constituent phases; olivine - orthopyroxene - clinopyroxene - spinel. The absence of water-bearing or hydroxyl minerals implies dry condition during the formation of the xenoliths. Without the present of garnet, and the coexisting of plagioclase and spinel in a few samples suggest that thinning lithosphere is beneath the study area, where the shallowest crust – mantle boundary may be at the depth approximately 33 km. In addition, the absent of garnet from the mineral assemblages reflects that the upper limit of the pressure should not exceed 25 kbar or the depth of 82 km, which is the beginning region of garnet lherzolite stability field (O'Hara, et al., 1971, O'Neill, 1981, Wilson, 1989)

The variation of rock types range from spinel lherzolite, harzburgite, and dunite indicates the evolution of these mantle peridotites as a remnant after several episodes of magma segregation, in addition, it also implies that the upper mantle beneath the study area is anomalous, and may be partly composed of spinel-plagioclase lherzolite. The host alkali basalt has Mg#, Mg/(Mg + Fe), 0.61-0.65, which is out of the range of Mg# that is in equilibrium with mantle olivine (Wilson, 1989), therefore, it is unlikely that these peridotites are cognate materials related to the host alkali basalt.

6.2 Geotherms beneath the Bo Ploi Area

The Bo Ploi palaeogeotherm from the maximum P-T estimate is, then, compared with mantle xenoliths from other localities as in figure 3. The majority of them lie in between the oceanic geotherm and the alkali province geotherm. The Bo Ploi xenoliths are clearly from higher pressure, implying deeper seated, when compare to xenoliths from Engeln, Lashaine, and Delegate, though these xenoliths are located along higher geothermal gradient.

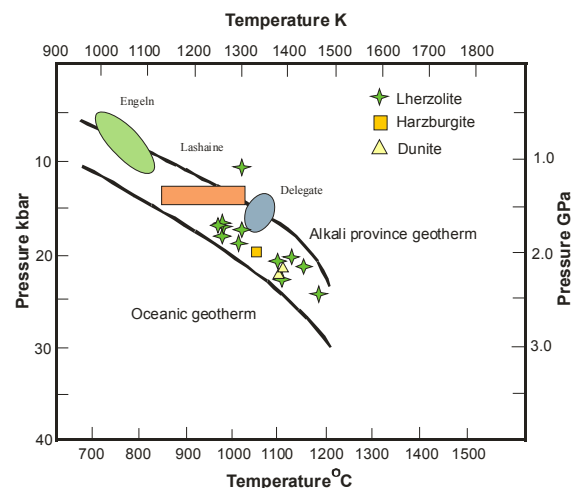


Figure 3. Thermodynamic calculations for the pressure and temperature on mineral pairs; enstatite-spinel (Mercier, 1980). The diagram shows the fields of lower crustal granulite xenoliths (Engeln, Lashaine, and Delegate) and alkaline province geotherm from Jones *et al.*, 1983 and reference there in.

In terms of heat flow in the area, the Bo Ploi estimated palaeogeotherm forms a parallel trend slightly lower than 90 mWm^{-2} (Fig.4), which is consistent to the

record of present heat flow in the region measured from drilled hole in the nearby Tertiary basin, Suphanburi basin, is 70 mWm^{-2} (Gonecome, personal communication). In addition, it intersects Eastern China geotherm at approximately 17 kbar and Fukue-Jima geotherm at greater depth.

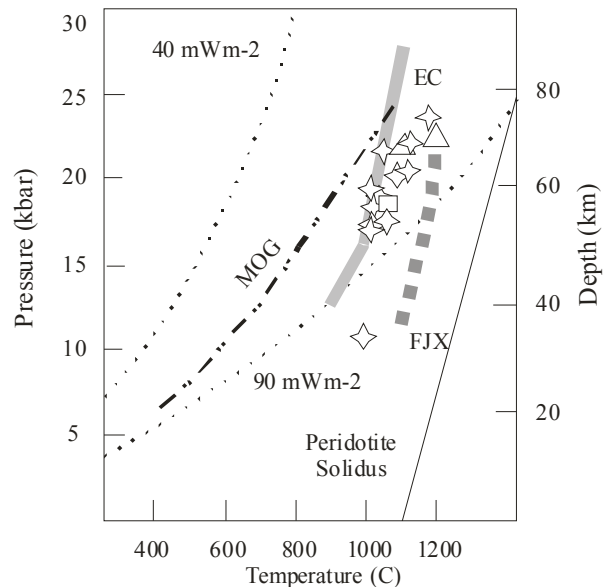


Figure 4 P-T estimation for Bo Ploi xenoliths (modified after Unimo & Yoshizawa, 1996, Nasir & Safarjalani, 2000, Al-Mishwat & Nasir, 2004), compare with those from Eastern China (EC; Fan&Hooper, 1989), Fukue-Jima, Japan (FJX; Unimo & Yoshizawa, 1996).

The occurrences of mantle xenoliths in alkali basalt - basanite - nephelinite magma series are usually associated with active tectonic setting, unlike mantle nodules found in kimberlite and its derivatives that are normally occurred in stable tectonic regions (Harte & Hawkesworth, 1989). The relatively high geotherm in this area, which is directly recorded from the geophysical measurement as well as the estimates from mantle xenoliths, imply the active tectonic regime, thus, interpreted as a result from thermal perturbation either by a regional intrusion of hot asthenosphere resulting from crustal extension or from deep-mantle plume. To date, the eruption of Late Cenozoic basalt in East Asia and Southeast Asia is still a controversy issue (Flower, *et al.*, 1998, Hoang and Flower, 1998, Jinfu *et al.*, 1998, Smith, 1998).

The crustal extension, which is a substantial reason for lithosphere thinning, is believed to play a role in the replacement of the continental lithosphere by the upwelling hot asthenosphere, which in turn elevates the geothermal gradient (Nicolas *et al.*, 1987, Witt and Seck, 1987, Umino and Yoshizawa, 1996). The Bo Ploi basalt is located along the major lineation structure and the eruption of may be triggered by the movement of two major fault zones in the vicinity, the Sri Sawat Fault Zone and the Three Pagoda Fault Zone, which have started in Mesozoic era with the sense of movement to the left and later transform to dextral movement in Later Cretaceous to Early Tertiary (Tapponnier *et al.*, 1986, Hall, 1996). The movement of these fault zones is a consequent event of the opening of the Gulf of Thailand as well as others Tertiary

basins, which are mostly N/S axes, after the collision-induced 'extrusion' of the East and Southeast Asia.

In conclusion, the entrained mantle xenoliths from Bo Ploi basalt are spinel lherzolite, dunite, and harzburgite. Their textures and mineral chemistry indicate mantle origin, though various degree of deformation may observe among these xenoliths. The estimate *P-T* scheme of the Bo Ploi xenoliths extends from 11 to 23 kbar corresponds to the depth of 33 to 78 km, which is in the upper mantle region and also stability field of spinel lherzolite. The coexisting of plagioclase and spinel in some xenoliths suggests that the influx of asthenosphere to the spinel – plagioclase transformation region and were later brought to the surface via the ascending alkali basalt. The estimated geotherm, which is agreed with the present day heat flow of 70 mWm⁻², reflects the high heat flow in the region.

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