煌斑岩的分类、特征及成因

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内容提要:作为最早被识别出的碱性岩石之一,煌斑岩因富含金和金刚石等矿产资源以及对理解深部地球动 力学过程的重要作用而受到广泛重视,但是目前对于煌斑岩的成因还存在不同的认识。本文基于近年来对煌斑岩 的研究成果,对它们的分类、特征以及岩石成因进行综述。根据国际地科联(IUGS)的分类标准,煌斑岩可以分为 超镁铁质煌斑岩、钙碱性煌斑岩和碱性煌斑岩。研究发现,超镁铁质煌斑岩往往是伸展环境下岩浆作用的产物,与 金伯利岩和碳酸岩有密切的成因关系;钙碱性煌斑岩通常发育在汇聚或被动大陆边缘环境,其岩石成因可能有多 种机制(如基性岩浆的分异、岩浆混合以及交代富集地幔的部分熔融);碱性煌斑岩出露在离散型大陆边缘和板内 构造环境,通常和碱性玄武质岩浆作用密切相关。不管岩石的形成环境和过程如何,超镁铁质煌斑岩、钙碱性煌斑 岩和碱性煌斑岩被普遍认为是来自于经历了交代富集作用的地幔源区。最后,文章指出了煌斑岩研究过程中存在 的一些科学问题,如富集的地幔源区存在的矿物相(金云母和角闪石)对产生钠质还是钾质岩浆的影响,控制岩浆 在结晶过程中影响含水矿物斑晶形成的因素以及部分煌斑岩中碳酸岩球粒和钠长石的形成原因等。

关键词: 煌斑岩;分类;岩石学特征;岩石成因;地幔交代作用

Von Gümbel 在 1874 年提出"煌斑岩 (lamporphyre)"一词,其词根来源于希腊语 "lampros porphyros",意思是闪闪发光的斑岩 (glistening porphyry)。研究初期,由于煌斑岩易与 其他斑状结构的火成岩(如玄武岩和安山岩)混淆, 一度被认为是"不可能分类(impossible to classify)"的火成岩,因此,早期煌斑岩的研究主要 集中在分类和命名方面。上世纪八九十年代,澳大 利亚西澳大学的 Rock 教授通过系统的研究后认 为,煌斑岩是一种特殊且十分重要的岩石类型,并发 表了一系列关于这类岩石的论文和专著(Rock, 1977, 1984, 1986, 1987, 1991),这些成果为之后 煌斑岩的深入研究奠定了重要的基础。

作为一种典型富含挥发分的镁铁一超镁铁质岩石,煌斑岩与金伯利岩以及碱性玄武岩等深部幔源 岩石一样,基本上只出露在大陆。部分煌斑岩因含 有金刚石或者与中温热液型金矿床密切相关 (Wyman et al., 2006; Muller and Groves, 2019) 而具有一定的经济价值,受到了业界的广泛关注。 煌斑岩岩浆具有较高的不相容元素含量,可在一定 程度上抵消地壳同化混染的影响,另外,煌斑岩中经 常包含一些地幔和地壳的深源捕掳体,这些捕掳体 对了解壳幔的结构和热状态也具有重要的指示意义 (Dostal et al., 2005; Orejana et al., 2006)。因此 煌斑岩对研究大陆岩石圈及软流圈地幔地球化学性 质以及地球深部动力学过程具有重要的作用(e.g., Guo et al., 2004; Tappe et al., 2006; Krmiček et al., 2011, 2016; Abdelfadil et al., 2013; Ulrych et al., 2014; Pandey et al., 2017a)。

煌斑岩可以分为不同的类型,包括超镁铁质煌 斑岩,钙碱性煌斑岩和碱性煌斑岩等(Rock,1991), 对于这些煌斑岩的成因和源区一直存在争议,目前 还没有一个统一的模型来解释煌斑岩诸多独特的岩 石学和地球化学特征(Currie and Williams, 1993; Le Roex and Lanyon, 1998; Riley et al., 2003; Prelević et al., 2004; Orejana et al., 2008; Tappe

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et al., 2008; Kerr et al., 2010; Soder et al., 2016; Maria et al., 2019)。但是,不管哪种类型的 煌斑岩以及它们的形成过程如何,都与经历过交代 富集的地幔源区密切相关。鉴于目前国内尚没有对 煌斑岩的研究现状进行系统的总结,本文将分别对 煌斑岩的分类、特征和岩石成因等方面进行综述,同时对煌斑岩研究过程中存在的一些问题进行探讨, 旨在为今后进一步研究煌斑岩的成因及其深部地质 过程提供重要信息。

1 煌斑岩的定义和分类

1.1 煌斑岩的定义和特征

煌斑岩是一系列富含挥发分(如 H_2O 和 CO_2) 的超基性--中性浅成岩,通常以岩脉的形式出露。 具有明显的斑状结构,斑晶可以由自形的橄榄石、辉 石、角闪石和黑云母等镁铁质矿物组成。其中,角闪 石和黑云母是最常见的斑晶(Rock, 1991)。该定义 是由 Rock 在 1991 年总结百年来地质学界对煌斑 岩的研究之后重新定义的,被大家广泛接受。早期 岩石学家将煌斑岩的研究重点集中在分类和命名上 面,由于不同种类的煌斑岩具有不同的矿物组合,而 且即使矿物组合相同,对应的矿物含量也千差万别, 因此对于煌斑岩的分类一直是一个棘手的问题。同 时,根据单纯的地球化学特征很难对煌斑岩与其他 正常的火成岩进行区分。由于煌斑岩中长石的比例 范围很不确定并且只会在基质中出现,而岩石本身 强烈的自交代作用又会导致长石发生蚀变,传统的 长石成分和比例的分类方式(如 QAPF 图解)也不 适合煌斑岩。国际地科联(IUGS)推荐的 Le Maitre (2002)主编的火成岩一书中总结了几种将煌斑岩与 其他岩石区分开来的方法:① 煌斑岩通常以岩脉的 形式出现:② 斑状结构,色率(M)一般在 35~90 之 间,呈中色至暗色,极少出现全暗色的煌斑岩,即 M >90 的煌斑岩:③ 长石或者长石族矿物只会出现 在基质中,不会以斑晶的形式出现;④ 肯定含有角 闪石或者黑云母(或富铁的金云母);⑤ 橄榄石、辉 石、云母以及长石的热液蚀变在煌斑岩中十分常见; ⑥ 方解石、沸石等矿物往往可作为原生矿物出现在 煌斑岩中;⑦ 与其他地球化学成分类似的岩石相 比,煌斑岩往往具有较高的 K₂O₂Na₂O₂H₂O₂CO₂、 S、P₂O₅和 Ba、Rb、Sr 含量。

1.2 煌斑岩的分类

Streckeisen (1979)利用矿物学和地球化学特征对煌斑岩进行了划分,将煌斑岩划分为:① 钙碱

性煌斑岩;②碱性煌斑岩;③黄长石质煌斑岩。这 种分类方法的主要依据是矿物组合,尤其是斜长石、 碱性长石和似长石的含量,并且被学者们广泛使用 (Rock, 1991; Mitchell, 1994a, 1994b; Le Maitre, 2002)。同时, Streckeisen(1979)还认为色率可以作 为辅助方法对以上三种煌斑岩进行区分。其中,钙 碱性煌斑岩的色率在 35~67 之间, 少部分在 25~ 35 之间;碱性煌斑岩的色率为 40~70 之间,黄长石 质煌斑岩的色率大于70。但是可以看出,钙碱性煌 斑岩和碱性煌斑岩的色率范围基本重合,所以色率 值只能作为煌斑岩的特征,而不是区分依据。Rock (1991)在 Streckeisen(1979)的研究基础上将煌斑 岩划分为五个分支,分别为:① 钙碱性煌斑岩(Calalkaline lamprophyres, CAL); ② 碱性煌斑岩 (Alkaline lamprophyre, AL);③ 超镁铁质煌斑岩 (Ultramafic lamprophyres, UML);④ 金伯利岩 (Kimberlites, KIL);⑤ 钾镁煌斑岩(Lamproites, LL)(表1)。需要注意的是,关于金伯利岩和钾镁煌 斑岩是否属于煌斑岩大类,一直存在争议。Rock (1991)认为可以将金伯利岩划分到煌斑岩大类是因 为金伯利岩几乎具有煌斑岩定义的全部特征,在研 究过程中两者往往会混淆(尤其是与超镁铁煌斑岩 如方解霞黄煌岩)。对于钾镁煌斑岩的识别也出现 了同样的问题,主要是由于其矿物组合和岩相学特 征与普通的煌斑岩十分相似。目前,很多学者都遵 循国际地科联认可的 Le Maitre(2002)划分标准(表 2)

值得说明的是,对于超镁铁质煌斑岩而言,将它 们纳入煌斑岩类经历了一个漫长的过程。Kranck (1939)将北美洲拉布拉多半岛艾利克湾(Labrador Aillik Bay)广泛出露的一种富碳酸盐的煌斑岩称为 方解霞黄煌岩(aillikite)。方解霞黄煌岩与之前发 现的另外一种超镁铁质硅不饱和岩石黄长煌斑岩 (alnoite)最大的区别就是缺乏黄长石。但是关于方 解霞黄煌岩的定名与发现并没有引起科学界的重 视。在早期版本中的 IUGS 岩石分类标准中只是简 单的提到了黄长石质煌斑岩,如黄长煌斑岩和橄黄 煌岩(polzenite)(Streckeisen, 1979)。Rock(1986) 正式提出了超镁铁质煌斑岩这一专业名词,并将超 镁铁质煌斑岩分为含黄长石和富含碳酸盐且不含黄 长石两种类型,即含黄长石的黄长煌斑岩和橄黄煌 岩,以及富含碳酸盐且不含黄长石的方解霞黄煌岩、 黑云碱煌岩(damkjernite)和黑云沸煌岩 (ouachitite)。尽管超镁铁质煌斑岩的提出对煌斑 岩的分类以及区分金伯利岩和煌斑岩具有重要的指示作用,但是最初并没有被 IUGS 所接受(Le Maitre, 1989, 2002; Woolley et al., 1996)。之后方解霞黄煌岩首次被 IUGS 采用,但被定义为是一种硅质碳酸岩,这一划分标准其实并不符合碳酸岩

的定义,因为碳酸岩要求岩石中的碳酸盐矿物比例 大于 50%(Le Maitre, 2002)。直到 2005年,IUGS 正式将超镁铁质煌斑岩作为煌斑岩的一种加入到火 成岩分类当中(Tappe, 2005),并且继续沿用含黄长 石和富含碳酸盐且不含黄长石两种类型。

表 1 Rock(1991) 煌斑岩分类

Fable 1	Classification	of	lamprophyres	(Rock,	1991)
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煌斑岩(Lamprophyric rocks)							
类型	钙碱性煌斑岩	碱性煌斑岩	超镁铁质煌斑岩	金伯利岩	钾镁煌斑岩		
简写	CAL	AL	UML	KIL	LL		
主要岩石类型	云煌岩 (minette) 云斜煌岩 (kersantite) 闪斜煌岩 (spessartite) 闪辉正煌岩 (vogesite)	闪煌岩 (camptonite) 沸煌岩 (monchiquite) 棕闪煌岩 (sannaite)	方解霞黄煌岩 (aillikite) 黄长煌斑岩 (alnoite) 黑云碱煌岩 (damtjernite)	Group I Group II	_		

表 2	IUGS 推荐的煌斑岩矿物学分类和命名表 (Le Maitre, 2	2002)
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Table 2 Mineralogical classification and nomenclature of lamprophyre recommended by IUGS (Le maitre, 2002)

浅色矿物		主要的镁铁质矿物			
长石	似长石	黑云母>角闪石, 土单斜辉石,(土橄榄石)	角闪石,单斜辉石, 土橄榄石	棕色角闪石,单斜辉石, 橄榄石,黑云母	
or>pl	_	云煌岩	闪正煌岩	_	
pl>or	_	云斜煌岩	闪斜煌岩	_	
or>pl	feld>foid	_	_	棕闪煌岩	
pl>or	feld>foid	_	_	闪煌岩	
	玻璃或 foid	_	_	沸煌岩	

注: or一碱性长石; pl一斜长石; feld一长石; foid 一似长石。

随着对煌斑岩地球化学和矿物学研究的进一步 深入,Rock(1987)认为可以利用全岩主量元素含量 对不同种类的煌斑岩进行简单的区分(图 1,图 2)。 同时,他对煌斑岩中矿物的研究发现,虽然矿物组合 相同,但是不同类型煌斑岩中角闪石、黑云母、长石 以及辉石的成分存在一定的差异,因此也可以利用 这些矿物的元素含量来区分不同种类的煌斑岩 (Rock, 1987)。同样的, Wilson(1989) 根据全岩的 Al₂O₃、K₂O以及 MgO 的含量对这三种类型的岩石 进行简单的区分(图 3)。从图 3 看出,金伯利岩具 有相对较高的 MgO 含量,钾镁煌斑岩具有相对较 高的 K₂O 含量。但是我们从详解判别图中可以看 出不同亚类煌斑岩所覆盖的区域有大量的重叠(图 1,图 2 和 Rock(1991)中的图 3),这使得我们很难 利用该图解精确地确定煌斑岩亚类。我们搜集了近 几十年来发表的不同类型煌斑岩数据,并且将这些 数据投入 Rock(1987)的煌斑岩判别图中发现,虽然 整体趋势一致,但是还是与主量元素判别图存在一 定的差异(图1,图2)。因此在区分这类富挥发分岩 石的过程中,不能单一的只利用全岩地球化学成分, 而是将矿物组合、矿物成分等多方面结合起来确定 其所属的种类。

考虑到国际地科联推荐分类方案以及金伯利岩 和钾镁煌斑岩本身特殊的研究意义,本篇综述只对 不具有争议的煌斑岩类型(钙碱性煌斑岩、碱性煌 斑岩和超镁铁质煌斑岩)的矿物组合和全岩地球化 学特征以及各类型所包含的煌斑岩种类进行详细的 总结和讨论。

2 煌斑岩的岩石学特征和鉴别

2.1 超镁铁质煌斑岩

鉴别超镁铁质煌斑岩首先是将其与金伯利岩和 钾镁煌斑岩区分开来。当一种超镁铁质岩石的色率 超过 90%,并且含有橄榄石+金云母巨晶/斑晶,则 要根据以下步骤鉴别其是金伯利岩、钾镁煌斑岩还 是超镁铁质煌斑岩:① 如果该岩石不含有原生碳酸



图 1 煌斑岩全岩组分分类(底图据 Rock, 1987,其中搜集的岩石数据按照 IUGS 煌斑岩分类标准分类,即矿物组合) Fig. 1 Classification diagram of lamprophyres compositions(after Rock, 1987), in which lamprophyre data collected are classified according to IUGS lamprophyre classification standard, i. e. mineral association UML-超镁铁质煌斑岩;AL-碱性煌斑岩;CAL-钙碱性煌斑岩;LL-钾镁煌斑岩

UML-Ultramafic lamprophyres; AL-alkaline lamprophyres; CAL-calcalkaline lamprophyres; LL-lamproites



图 2 煌斑岩全岩组分分类三角图(底图据 Rock, 1987) Fig. 2 SiO₂/10-CaO-TiO₂×4 and Al₂O₃-MgO-CaO ternary diagrams (after Rock, 1987) for worldwide lamprophyres

盐矿物,参考钾镁煌斑岩的鉴别标准确定其是否为 钾镁煌斑岩;②如果含有原生碳酸盐矿物,并且含 有黄长石,则该岩石为黄长煌斑岩;③如果不含黄 长石而基质中含有霞石/碱性长石,该岩石为黑云碱 煌岩;④如果既不含黄长石也不含似长石和碱性长 石,而富碳酸盐并且含有富 Ti 石榴子石(黑榴石/钛 榴石或钙锆榴石),该岩石为方解霞黄煌岩(图 4)。 如果该岩石不具有以上描述的特征,则要根据尖晶 石、金云母或者单斜辉石的化学成分来判别。金伯 利岩中的金云母具有富 Ba-Al 的特征(钡镁脆云 母),并且基质中不会出现单斜辉石;而 orangeite (金伯利岩的一种)和方解霞黄煌岩中的金云母往往 具有富 Ti 贫 Al 的特征,同时 orangeite 中的尖晶石 Cr* 值[=100×Cr/(Cr+Al)]往往大于 85,方解霞 黄煌岩中的尖晶石 Cr* 往往小于 85 (Tappe et al., 2005)。从以上的区分方法可以看出,超镁铁质煌斑





Fig. 3 Al₂O₃-K₂O-MgO geochemical classification diagram of lamprophyres, lamproites and kimberlites (after Wilson, 1989)

岩中最常见且最主要的三种类型为黄长煌斑岩,方 解霞黄煌岩和黑云碱煌斑岩(表 3,图 5a, b),其主 要特征如下:

黄长煌斑岩是一种含黄长石的超镁铁质煌斑 岩,主要特征为含橄榄石,金云母和单斜辉石斑晶或 者巨晶。基质由黄长石、单斜辉石、金云母、尖晶石、 钛铁矿、钙钛矿、富钛石榴子石、磷灰石和少量原生 碳酸盐组成,极少情况下会出现蒙脱石。

方解霞黄煌岩是一种富含碳酸盐的超镁铁质煌 斑岩,其特征为橄榄石和金云母斑晶或者巨晶,基质 主要有原生碳酸盐、金云母、尖晶石、钛铁矿、金红 石、钙钛矿、富钛石榴子石和磷灰石。发生过交代作 用的方解霞黄煌岩由于基质中的碳酸盐被单斜辉石 或者角闪石所替代,色率往往>90%。

表 3 根据特征矿物超镁铁质煌斑岩分类(Tappe, 2005) Table 3 Classification of ultramafic lamprophyres based

on	their	diagnostic	mineralogy	(Tappe,	2005)	ł
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	黄长石 (madara)	碳酸盐 (and m	霞石 (madana)	碱性长石 (andra)
	(gram.)	(gram., 原生)	(gram.)	(gram.)
黄长煌斑岩	MN	m	m	—
方解霞黄煌岩	_	MN	_	_
黑云碱煌岩	_	m	mN *	mN*

注:以上三种类型超镁铁质煌斑岩中橄榄石和金云母都很常见,也 可能含有富 Ti 石榴子石,所以不作为判断依据。其中,grdm一基 质; M—主要组分; m—次要组分; N—必要矿物; N*—必须出现 其中的一种矿物即可(即基质中霞石和碱性长石至少出现其中一 种);具体判别流程见图 4。 黑云碱煌岩是含似长石或碱性长石的超镁铁质 煌斑岩,斑晶主要有橄榄石、金云母和单斜辉石。基 质由金云母/黑云母、单斜辉石、尖晶石、钛铁矿、金 红石、钙钛矿、富钛石榴子石、磷灰石、原生碳酸盐以 及必需的霞石和/或碱长石组成。

2.2 钙碱性煌斑岩

钙碱性煌斑岩的色率在 35~67 之间,主要由原 生钾长石(Or₅₀₋₉₈)、斜长石(钙长石到钠长石)、黑云 母/金云母和角闪石组成,也可能含有橄榄石 (Fo₈₀₋₉₁)、贫Ti透辉石(基质也可能出现霓石)、磷灰 石、榍石、绿帘石以及原生的碳酸盐和沸石。但是不 会出现白榴石、黄长石、白云母、霞石、斜方辉石、钙 钛矿、易变辉石、方钠石等矿物。Streckeisen (1979)根据所含长石和暗色矿物的种类,将钙碱性 煌斑岩分为以下几种:云煌岩(图 5c, d)、云斜煌岩、 闪正煌岩、闪斜煌岩、钠云煌岩等(Rock, 1984)(具 体详见表 4)。

表 4 根据特征矿物的钙碱性煌斑岩分类(Streckeisen, 1979) Table 4 Rock-type definitions of calc-alkaline lamprophyre based on Streckeisen (1979)

		1	
主要长石矿物		主要镁铁质矿物	1
	黑云母-金云母	角闪石	钠质辉石-角闪

	黑云母-金云母	角闪石	钠质辉石-角闪石
钾长石	云煌岩	闪正煌岩	钠云煌岩
斜长石	云斜煌岩	闪斜煌岩	_

钙碱性煌斑岩具有以下典型的岩相学特征:① 黑云母斑晶具有环带结构(暗色的边部和浅色的核 部);② 含有磷灰石斑晶;③ 斑晶中不含长英质矿 物并且基质中不会出现橄榄石;④ 以浅色矿物为主 的"球粒结构"比较常见,球粒主要由碳酸盐、绿泥 石、绿帘石、低温长石和石英组成,偶尔会有硫酸盐 矿物出现。与闪煌岩中的"球粒"结构明显不同(详 见碱性煌斑岩球粒结构具体描述)。如果忽略其化 学成分,单从矿物学的角度对钙碱性煌斑岩与碱性 煌斑岩进行区分,是要根据基质中是否出现似长石 矿物,如霞石,方钠石等。同时,钙碱性煌斑岩接近 于硅饱和(标准矿物含量<15%石英,<10%霞石), 并且大多数属于过铝质。

2.3 碱性煌斑岩

由于与钠质碱性岩石具有成因上的联系,比如 许多闪煌岩和沸煌岩常常与霞石正长岩/辉长岩岩 体共存,因此将这些常与碱性岩石共存的煌斑岩称 为碱性煌斑岩(Rosenbusch, 1887)。Wimmenauer (1973)认为这些碱性煌斑岩与玄武岩的成分相似, 称之为"近玄武质"。之后 Wimmenauer 进一步发







Fig. 4 Flow chart illustrating how to distinguish between the three UML end-members(alnoite, damtjernite, aillikite), orangeite and kimberlite (after Tappe, 2005)

现去除这些碱性煌斑岩较高的挥发分之后其成分与 碱性玄武岩或碧玄岩相同,故称之为碱性玄武质或 碧玄质(Rock,1977)。Rock(1977)通过对比碱性 玄武岩和碧玄岩与上百个碱性煌斑岩的样品的主量 元素发现,碱性煌斑岩与碱性玄武岩/碧玄岩的成 分几乎相同,如果要将它们区分开来,只能根据矿 物组成,比如是否具有角闪石或者云母的斑晶。 碱性煌斑岩的色率一般在40~70之间。其中,闪 煌岩、沸煌岩、棕闪煌岩为碱性煌斑岩最主要的三 种类型,区别这三种碱性煌斑岩类型主要是看基 质中的矿物种类和相对含量。闪煌岩基质中斜长 石占主导地位,碱性长石/似长石含量较少,而棕 闪煌岩则正好相反,是碱性长石/似长石占主要地 位,沸煌岩的基质往往是玻璃质,或者由方沸石或 者霞石组成:

闪煌岩:斑晶主要由角闪石(铁角闪石、钛角闪石)、单斜辉石、橄榄石和黑云母组成,基质中含有斜长石,角闪石,单斜辉石以及少量的碱性长石或似长石,也可能含有磷灰石、金属氧化物、方解石和沸石等矿物(图 5e, f)。

棕闪煌岩:矿物组成与闪煌岩相似,但是基质中 碱性长石的含量高于斜长石。

沸煌岩:斑晶主要由单斜辉石、角闪石(铁角闪石、钛角闪石)、黑云母组成,基质往往是玻璃质,或 者是由方沸石或霞石组成。其中,无橄沸煌岩 (fourchite)是经常在文献中出现的沸煌岩类型,它 代表一种无橄榄石的沸煌岩。

Rock(1977)认为可以根据煌斑岩中实际矿物



图 5 煌斑岩镜下照片

Fig. 5 Photomicrographs of the lamprophyres

(a-b)一超镁铁质煌斑岩,单偏光(图片引自 Pandey et al., 2017c);(c-d)一钙碱性煌斑岩(云煌岩),单偏光(图片引自 Soder et al., 2018);
 (e-f)一碱性煌斑岩(闪煌岩),单偏光(图片引自 Van der Meer et al., 2016);Amp一角闪石;Cpx一单斜辉石;Foid一碱性长石;Ol一橄榄石;
 Op一斜方辉石;Phl一金云母;Pv-钙钛矿;Tm-钛磁铁矿;Kr-钛闪石

(a-b)—Ultramafic lamprophyres, plane-polarized light (after Pandey et al., 2017c); (c-d)—calc-alkaline lamprophyres (minette), planepolarized light (after Soder et al., 2018) and (e-f)—calc-alkaline lamprophyres (camptonite), plane-polarized light (after Van der Meer et al., 2016); Amp—amphibole; Cpx—clinopyroxene; Foid—feldspathoid; Ol—olivine; Op—orthopyroxene; Phl—phlogopite; Pv perovskite; Tm—titanomagnetite; Kr—kaersutite

的含量利用 QAPF 双三角图解对以上描述的几种 煌斑岩类型和各类型所包含的煌斑岩种类进行划分 (见 Rock, 1977 中图 3)。需要注意的是,上述描述 的区分煌斑岩类型的方法大多和浅色矿物有关(如 碱性长石、斜长石、似长石),但是煌斑岩中的浅色矿 物基本上只会出现在基质中,粒度很细并且容易发 生蚀变,需要十分细致的观察和测试才能区分。因此在确定岩石为煌斑岩但又不确定具体是哪种类型,可以借助现在先进的仪器对岩石中的矿物进行 定量和定性扫描(如 TIMA),然后确定其矿物组合, 这样来确定煌斑岩的具体类型。

3 煌斑岩的岩石成因

前已述及,根据 IUGS 的分类标准,煌斑岩分为 超镁铁质煌斑岩、钙碱性煌斑岩和碱性煌斑岩三种 类型,下面将分别对这三种类型煌斑岩的成因进行 评述。

3.1 超镁铁质煌斑岩

由于超镁铁质煌斑岩在岩石特征上与金伯利岩 存在一定的相似性,暗示它们在成因上也可能存在 某种共性。另外,超镁铁质煌斑岩通常和碳酸岩共 生在裂谷边缘形成岩墙群或杂岩体(Tappe et al., 2006),说明其与碳酸岩也存在着密切的成因联系。 下面将分别讨论超镁铁质煌斑岩与金伯利岩和碳酸 岩的成因关系。

3.1.1 超镁铁质煌斑岩与金伯利岩的关系

关于超镁铁质煌斑岩的岩石学成因还存在诸多 争议,由于富含碳酸盐的超镁铁质煌斑岩(方解霞黄 煌岩和黑云碱煌岩)通常与金伯利岩密切相关 (Tappe et al., 2005)。所以争论的核心是方解霞 黄煌岩和金伯利岩之间的成因关系,以及原生超镁 铁质煌斑岩熔体产生过程中岩石圈所扮演的角色, 其中包括岩石圈组分所贡献的量(Foley et al., 2009; Francis and Patterson 2009; Tappe et al., 2006; Mitchell and Tappe 2010; Veter et al., 2017)。金伯利岩的源区通常被认为是大陆岩石圈 底部或者之下的软流圈地幔,根据计算金伯利岩所 携带的最深的地幔捕掳体平衡时的温压以及氧化还 原条件为1350℃、7GPa,处于金刚石和CH4的稳定 域(Malkovets et al., 2007)。超镁铁质煌斑岩岩浆 中也可能含有金刚石(Hamilton, 1992; Mitchell et al., 1999; Digonnet et al., 2000; Birkett et al., 2004),这指示了形成超镁铁质煌斑岩岩浆的熔融深 度可能超过 150km(压力大于 5 GPa),同时也说明 熔融深度可能并不是金伯利岩和超镁铁质煌斑岩最 关键的岩石成因差异。Francis and Patterson (2009)认为方解霞黄煌岩和金伯利岩在成因上有一 定的联系。其中,碳酸盐化二辉橄榄岩在 8~10GPa 的条件下发生低程度部分熔融可以形成富碳酸盐熔 体,在克拉通环境下,这些富集碳酸盐的熔体可能会 与大陆克拉通之下的岩石圈地幔中高 Mg 方辉橄榄 岩反应,导致熔体中 SiO₂的升高,形成金伯利岩岩 浆,而在非克拉通的地质背景下由于软流圈的 Mg 含量较低,这种富集碳酸盐的熔体发生演化形成方 解霞黄煌岩岩浆。Francis and Patterson(2009)认 为方解霞黄煌岩熔体不能穿过厚的岩石圈地幔发生 侵位形成方解霞黄煌岩,这个过程中方解霞黄煌岩 熔体即使不与岩石圈地幔发生反应也会发生演化转 变为玄武质熔体。但是他们忽视了自然界中确实存 在的方解霞黄煌岩岩浆穿过厚的克拉通岩石圈的证 据,格陵兰岛以及加拿大拉布拉多地区的超镁铁质 煌斑岩(方解霞黄煌岩)岩浆就是穿过了厚的克拉通 岩石圈地幔发生侵位的(Foley, 1989; Tappe et al., 2006),并不是他们在假设中描述的那样形成 金伯利岩或玄武质岩石。大多数观点认为方解霞黄 煌岩以及其他超镁铁质煌斑岩是地幔橄榄岩与含金 云母交代脉混合发生熔融的产物,并且认为金伯利 岩和方解霞黄煌岩之间并不存在成因上的联系 (Foley et al., 2002, 2008; Tappe et al., 2006, 2007, 2008; Upton et al., 2006; Mitchell and Tappe, 2010; Nasir et al., 2016; Nosova et al., 2018)。在该成因模型中,软流圈或地幔柱产生的少 量原生富碳酸盐熔体可以在克拉通岩石圈底部冷凝 形成碳酸盐-金云母脉,这些脉体在后期热的作用下 发生熔融形成富挥发分的熔体,产生的熔体会与周 围石榴子石橄榄岩反应,形成原生方解霞黄煌岩熔 体(Tappe et al., 2006, 2007)。总之,富含碳酸盐 的超镁铁质煌斑岩和金伯利岩不管岩石成因上有没 有相关性,但是由于两者的岩相学和地球化学特征 方面具有很高的相似性,这就说明两者在源区或者 其形成过程可能具有相似性,如源区的矿物组成或 者岩浆演化过程中的物理化学条件。

3.1.2 超镁铁质煌斑岩与碳酸岩的关系

自 Brøgger(1921)首次提出岩浆成因的碳酸岩 之后,该岩石的岩石学成因一直是地质学界感兴趣 的问题。许多研究集中在各种与之共生的碱性硅酸 盐岩石类型在其形成过程中的作用,但是这些碱性 硅酸盐岩浆的多样性似乎就决定了碳酸岩岩浆形成 方式的多样性(Bell, 1998; Mitchell, 2005)。其 中,方解霞黄煌岩(或其他富碳酸盐的超镁铁质煌斑 岩)是最重要的与碳酸岩产出或成因有关的岩浆类 型,在碳酸岩的研究过程中起到十分重要的作用 (Beard et al., 1996; Tappe et al., 2006, 2009; Woolley and Kjarsgaard, 2008; Nasir et al., 2011; Smith et al., 2013; Krüger et al., 2013; Srivastava et al., 2013)。目前被广泛接受的碳酸 岩成因主要有两种:① 直接来自碳酸盐化地幔橄榄 岩的部分熔融(Dalton and Presnall, 1998; Mitchell et al., 1999; Mitchell, 2005; Woolley and

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Kjarsgaard, 2008);② 碳酸岩岩浆作为一种演化的 岩浆,可能是富碳酸盐的超镁铁质岩浆不混溶作用 形成,其产物为碳酸岩和方解霞黄煌岩岩浆(Tappe et al., 2006);③碳酸岩和方解霞黄煌岩可能来自 同一母岩浆,其中,碳酸岩是发生矿物分离结晶后堆 晶的产物,而残留的岩浆形成方解霞黄煌岩(Tappe et al., 2009; Smith et al., 2013, Krüger et al., 2013)。需要注意的是这几种成因模型要求产生碳 酸岩岩浆的深度(压力)可能不同,其中不混溶作用 和分离结晶作用似乎可以在较浅的深度进行,而碳 酸岩如果代表原始岩浆,这就要求形成的深度更大 (Kiseeva et al., 2012)。关于第一种成因模式, Dalton and Presnall(1998)根据实验岩石学模拟了 在 6GPa (金刚石稳定区) 碳酸盐化地幔橄榄岩发 生不同程度部分熔融并且熔融程度增加依次形成碳 酸岩,方解霞黄煌岩和金伯利岩。并且他们根据 CaO-MgO-Al₂O₃-SiO₂-CO₂体系与西格陵兰岛地区 出露的碳酸岩,方解霞黄煌岩和金伯利岩进行对比 发现,实验熔体与实际十分吻合,因此推测这种模型 适用于解决全球其他地区碳酸盐-金伯利岩-超镁铁 质碱性岩体系的成因问题。不混溶和分离结晶模式 的共同点就是母岩浆都可能是超镁铁质煌斑岩岩浆 (最初始的方解霞黄煌岩岩浆),主要根据共生的方 解霞黄煌岩和碳酸岩的侵位年龄以及矿物学和地球 化学特征认为碳酸岩岩浆是最初始的富含碳酸盐硅 酸盐岩浆演化的产物,证据就是方解霞黄煌岩具有 原始岩浆的矿物学特征,如高 Mg 橄榄石和富 Cr 尖 晶石微晶,而碳酸岩则显示出演化的特征,如富 Fe 的白云石(Tappe et al., 2009),并且在自然界中也 存在符合这两种模式的最初始的方解霞黄煌岩样品 (Tappe et al., 2006)。虽然第一种观点有实验岩 石学和自然样品作为证据支持,但是可能并不适用 于一些后来才发现的具有明显演化特征的碳酸岩体 系。而后二种观点正好弥补了第一种观点所忽略的 问题,同时又出现了另外一个亟待解决的问题,那就 是如果初始的方解霞黄煌岩岩浆代表了碳酸岩的母 岩浆,那么碳酸岩熔体如何从这种母岩浆分离出来 的?其不混溶机制和分离结晶的条件又是什么?

3.2 钙碱性煌斑岩

钙碱性煌斑岩通常发育在汇聚或被动大陆边缘 环境(Hoch et al., 2001; Owen, 2008; O'Leary et al., 2009; Garza et al., 2013; Karsli et al., 2014; Krmíček et al., 2014; Ma Liang et al., 2014),并 常常与钾玄质或钙碱性火成岩密切共生(Rock, 1991)。关于钙碱性煌斑岩的岩石成因,主要有 3 种 不同观点:① 基性岩浆的分异(Currie and Williams, 1993);② 岩浆混合(Prelević et al., 2004; Xu Xingwang et al., 2007; Ivanov et al., 2019);③ 交代富集岩石圈地幔低程度部分熔融 (Leat et al., 1988; Hoch et al., 2001; Chen et al., 2003; Guo et al., 2004; Chalapathi Rao et al., 2012; Abdelfadil et al., 2013; Xiong Fuhao et al., 2013; Garza et al., 2013; Imaoka et al., 2014; Karsli et al., 2014; Aghazadeh et al., 2015; Soder et al., 2016; Pandey et al., 2017b; Soder and Romer, 2018; Choi et al., 2019)。接下 来将分别介绍以上 3 种成因观点。

3.2.1 基性岩浆的分异

Currie and Williams(1993)根据区域上出露的 镁铁质以及长英质岩石与煌斑岩之间的密切关联推 断三者之间存在着演化的关系。在这个过程当中煌 斑岩岩浆是拉斑玄武质岩浆在上升过程中由于地壳 物理化学条件变化演化形成的。判断依据一是根据 煌斑岩具有低 MgO、Ni 和 Cr 含量,说明其是演化 的岩浆;二是根据煌斑岩与其他镁铁质岩浆的差别 (是否具有斜长石/含水矿物斑晶),推测前者是后者 在上升到地壳由于挥发分的加入演化形成的。作者 的目的是建立煌斑岩与拉斑玄武质-钙碱性岩浆系 列演化的关系,但是似乎缺乏关键的证据说明煌斑 岩与镁铁质岩浆之间的继承关系,比如矿物学证据; 如果是镁铁质岩浆上升过程受到了流体的加入才发 生演化形成煌斑岩岩浆,那么流体加入的证据是什 么?还有最重要的一点就是岩浆演化分异过程中都 是从超基性一基性一中性一酸性方向进行,基性岩 浆发生演化是否继续形成与它 SiO₂ 含量相近的煌 斑岩岩浆同样存在疑问。也许部分地区存在煌斑岩 与基性岩密切的演化关系(前提是煌斑岩的 SiO₂含 量要小于与其成因相关的基性岩 SiO2含量,这样才 行得通),但可能并不适用于大多数的钙碱性煌斑 岩。部分学者认为幔源的钙碱性煌斑岩岩浆在上升 的过程中会受到地壳物质的混染(O'Leary et al., 2009; Ma et al., 2014), 是因为会出现一些地壳混 染的地球化学信号(如 $\epsilon_{Nd}(t)$ 和²⁰⁷ Pb/²⁰⁴ Pb)。但这 并不意味着地壳混染影响到了本身来源于俯冲相关 的交代富集地幔的煌斑岩母岩浆整体性质(文章中 作者也有提到),幔源岩浆在上升过程中必然会受到 一定程度的地壳混染,因此没有必要单独将幔源基 性岩浆受到地壳的混染作为一种成因模式单独提出

(Chalapathi Rao et al., 2012; Pandey et al., 2017b),但是来自于地壳的熔体组分对幔源熔体的影响则另当别论(如下讨论)。

3.2.2 岩浆混合

Rock(1984, 1991)在对煌斑岩进行综合研究评 述时就提出,钙碱性煌斑岩可能是钾镁煌斑岩熔体 和壳源硅质熔体混合的产物。这是由于钙碱性煌斑 岩通常产出于造山带等汇聚型的构造环境,并且与 长英质的侵入体共生,而钾镁煌斑岩作为一种罕见 的富 K、富挥发分幔源岩石,在汇聚型的构造环境中 常常和钙碱性煌斑岩或其他钙碱性硅质岩浆活动密 切相关(Mitchell and Bergman, 1991)。因此 Rock (1991)根据钙碱性煌斑岩特有的矿物和地球化学特 征(钙碱性煌斑岩通常具有地壳组分混入的信号,并 且具有和钾镁煌斑岩相似的矿物组合)推测在造山 带内部,尤其是花岗质岩浆活动活跃的地方,幔源的 钾镁煌斑岩熔体很难穿过厚的大陆地壳,会在上升 的过程中极容易受到壳源组分的改造,从而变成钙 碱性煌斑岩(云煌岩)熔体。关于这种假设,Rock (1991)并没有提出具体的成因模型以及这个过程中 岩浆混合的具体过程,而且自然界中极少发现钾镁 煌斑岩-钙碱性煌斑岩-长英质岩石共生的例子。 Prelević et al. (2004)报道了塞尔维亚 Veliki Majdan 地区共生的钾镁煌斑岩-云煌岩-长英质岩 石杂岩体的矿物学和地球化学数据,并且找到了该 成因模型的具体证据,如黑云母的海绵状熔蚀结构, 部分金云母化和核部为黑云母而边部为金云母的反 环带(混合过程中长英质熔体的加入以及温度过高 导致的),以及主微量和同位素混合证据(SiO₂, Zr, Zr/Nb, ¹⁴³Nd/¹⁴⁴Nd_i)。这种岩浆混合模型可以解 释地球上其他造山环境中与钙碱性花岗质侵入体密 切相关云煌岩的岩石成因,同样,也可以解释钙碱性 煌斑岩时常出现的金云母反环带和熔蚀结构现象 (Prelevic et al., 2004 以及所引的相关参考文献)。 关于岩浆混合的模型还有以下几种:① Xu Xingwang et al. (2007)根据区域上出露的玄武岩和 碱性岩体认为云南北衙金矿区钙碱性煌斑岩是玄武 质熔体和富 SiO2碱性熔体混合的产物,其证据有透 辉石环带中 SiO₂ 含量的突然升高、矿物(透长石和 霓辉石)的不平衡反应边、碱性岩石的包裹体以及橄 榄石和石英的同时出现等; ② Ivanov et al. (2019) 根据不平衡的矿物组合(富 Mg 橄榄石和石英)推测 西伯利亚克拉通后碰撞环境钙碱性煌斑岩是岩浆混 合的产物,两个端元分别是俯冲地壳物质和交代富 集的岩石圈地幔发生熔融产生的熔体。综上所述, 关于岩浆混合的几种观点都是根据岩石中存在的混 合作用的证据,如单矿物的不平衡结构、不平衡的矿 物组合、区域上共存的混合端元的主微量元素以及 同位素地球化学等方面的证据。

3.2.3 交代富集地幔低程度部分熔融

鉴于钙碱性煌斑岩通常具有富集大离子亲石元 素(LILE)和轻稀土元素(LREE)以及亏损高场强元 素(HFSE)的特征,因此大多数研究认为钙碱性煌 斑岩岩浆是一个经历过交代的富集地幔源区发生低 程度部分熔融产生的,其中富集的地幔源区可能是 岩石圈地幔也可能是地幔楔区域,而富集组分可能 是来自于俯冲板片所释放的流体和/或熔体(Rock, 1984, 1991; Leat et al., 1988; Hoch et al., 2001; Garza et al., 2013; Xiong Fuhao et al., 2013; Karsli et al., 2014; Soder and Romer, 2018; Cho et al., 2019),也可能是来自于深部软流圈的交代 富集介质(Chen et al., 2003; Aghazadeh et al., 2015)。这些不同性质的交代介质渗透到地幔引发 富集作用的具体情况是比较复杂的,但是可以肯定 的是交代介质会与地幔岩石发生反应改变其矿物相 和/或地球化学组成,从而可能导致上地幔成分和矿 物的不均一性(Fumagalli and Klemme, 2015; O' Reilly and Griffin, 2013)。下面我们将对地幔交代 作用进行详细的阐述,以期将煌斑岩地球化学、矿物 学特征以及岩浆形成过程和地幔交代作用紧密的联 系起来。

3.2.4 地幔交代作用

经过长期的研究人们已经认识到大陆岩石圈地 幔在化学和矿物学成分上都是不均一的,正是这种 不均一性才保留了长期以来岩石圈地幔交代富集的 复杂记录(Hawkesworth et al., 1984, 1990a: Erlank et al., 1987; Waters and Erlank, 1988; Harte and Hawkesworth 1989)。通过详细的地幔 捕掳体研究,可以将地幔交代作用分为两种方式:第 一种称为显交代作用(modal metasomatism),是指 在交代作用过程中除了常见的橄榄岩矿物之外,还 会有角闪石、云母、碳酸盐、磷灰石、硫化物等含挥发 分矿物的出现。第二种为隐交代作用(cryptic metasomatism),是指在没有新矿物出现的情况下, 原生矿物主微量以及同位素地球化学方面的富集 (Dawson, 1984)。导致这两种交代作用发生必须 要有地幔岩石和交代介质的参与。与地幔发生交代 作用的介质包括硅酸盐熔体、碳酸岩熔体、硫化物熔 体、C-H-O 流体(CH₄和 CO₂)、卤水流体、含烃流体 和超临界流体。在许多情况下这些介质是可以共存 的(O'Reilly and Griffin, 2013)。参与以上反应的 岩石圈地幔主要岩石类型包括以下几种:① 地幔橄 榄岩,占岩石圈地幔总体积的 95%以上,属于超镁 铁质岩石,其中包括方辉橄榄岩,异剥橄榄岩和二辉 橄榄岩。这些橄榄岩地幔中还富 Al 的矿物相,如 石榴子石,尖晶石和斜长石,这三种矿物的存在受温 度压力的影响;② 榴辉岩为由单斜辉石和石榴子石 组成镁铁质岩石,属于地幔高压低温条件下的岩石 类型。它可能是洋壳变质的产物(Jacob, 2004),也 可能属于单纯的火成岩侵入体(Griffin and O' Reilly, 2007);③ 辉石岩成分上与玄武质岩石相 似,是分离结晶和堆晶的产物。辉石岩绝大多数侵 位于年轻的(显生宙)岩石圈地幔,很少出现在较老 的克拉通岩石圈地幔中(Downes, 2007)。

3.2.4.1 富水硅质熔体的地幔交代作用

地幔交代作用发生最显著的地方便是俯冲带。 洋壳主要由含水的镁铁质岩石(辉长岩和蚀变的大 洋中脊玄武岩)和沉积物组成(Ulmer, 2001; Stern, 2002),在俯冲过程中由于物理化学条件的 改变使得洋壳发生变质以及相转变(角闪岩相一榴 辉岩相转变,约2~2.5 GPa, Poli, 1993)。这个过 程会发生脱水反应或部分熔融,产生的流体和熔体 会与俯冲板片上方的地幔发生反应,从而改变地幔 的矿物和化学成分(Perchuk et al., 2013)。岩相学 和实验岩石学研究表明含水的榴辉岩相部分熔融产 生的富水富 Al 的硅质熔体形成类似 TTG 岩石组 合的高含水量和高 Al 的安山质到英安质的埃达克 岩(Jahn et al., 1981; Klemme et al., 2002; Gervasoni et al., 2017)。其中,板片发生熔融在太 古宙的俯冲系统中时常发生,因为那时的上地幔温 度高能使板片发生熔融(Rapp et al., 1991),这些 镁铁质的板片在榴辉岩或石榴子石角闪岩相发生熔 融会形成一套典型的 TTG 岩石组合(奥长花岗岩、 英云闪长岩、花岗闪长岩, Jahn et al., 1981)。而对 于现代的俯冲系统,只有年轻的热的洋壳岩石圈在 俯冲过程中才能发生熔融产生埃达克岩岩浆 (Peacock et al., 1994; Stern, 2002)。不管是太古 宙还是现代俯冲系统,这些富水硅质岩浆都会在上 升过程中消耗地幔橄榄岩中原生矿物相产生富 Al 斜方辉石和富 Mg 角闪石的双矿物相,反应方程式 为 Ol + Cpx + Opx(1) + Sp + 熔体 = Amp +Opx(2),其中形成富 Al 斜方辉石的反应方程式为 Al_2O_3 (熔体) + Mg₂ SiO₄ (Ol) = MgAlAlSiO₆ (Opx) + MgO (熔体) (Mallik and Dasgupta, 2012)。这一结果不仅得到了实验岩石学数据的支 持 (Sen and Dunn, 1994; Mallik and Dasgupta, 2012; Gervasoni et al., 2017),也在自然界岩石捕 掳体样品中找到了这种富集机制的证据(斜方辉石 和角闪石组成的交代脉,Grégoire et al., 2001, 2008; McInnes et al., 2001; Franz et al., 2002; Arai et al., 2003)。以上的交代反应不仅会在岩石 圈地幔中发生,在地幔楔中同样会产生斜方辉石和 角闪石组成的交代脉(Foley, 1992; Coltorti et al., 2007),这种交代富集的地幔楔发生部分熔融会产生 富水高 LILE 含量的钙碱性岩浆(Ulmer, 2001)。

3.2.4.2 富碳酸盐熔体的岩石圈地幔交代作用

俯冲过程的脱挥发作用(devolatilization)只能 消耗俯冲板片中的一小部分碳酸盐,剩余的大部分 碳酸盐会通过俯冲带进入深部地幔(Yaxley et al., 1994; Dasgupta et al., 2005)。进入地幔的碳酸盐 可能会储存在地幔中或者被抽取熔融形成碳酸岩或 富碳酸盐的熔体(Wallace and Green 1988; Dasgupta et al., 2005)。这些熔体由于具有低密度 和易迁移等特征,在地幔交代作用过程中扮演着十 分重要的角色(Blundy and Dalton, 2000; Klemme et al., 1995; Wallace and Green, 1988)。实验岩 石学的研究结果表明碳酸岩或富碳酸盐的熔体与地 幔橄榄岩发生反应会产生异剥橄榄质的矿物集合 体。这一结果在自然界的样品中也找到了相关证据 (Yaxley et al., 1998, 1991; Rudnick et al., 1993; Neumann et al., 2002)。上述提到的两种熔体(碳 酸岩熔体和富碳酸盐的硅质熔体)与地幔橄榄岩发 生反应都会形成橄榄石、单斜辉石和石榴子石。其 中,形成新的橄榄石和单斜辉石是消耗地幔橄榄岩 中的斜方辉石,其方程式为 2Mg₂ Si₂ O₆ (Opx) + $CaMg(CO_3)_2($ kpt $Mg_2SiO_4(Ol) + CaMgSi_2$ $O_6(Cpx) + CO_2(\hat{m} \Phi)$ (Wallace and Green, 1988; Dalton and Wood, 1993; Yaxley et al., 1998; Dasgupta and Hirschmann, 2007)。该反应过程中 的压力会影响形成单斜辉石中 Al₂O₃和 TiO₂的含 量,高压条件下形成的单斜辉石相比低压条件下含 有更低的 Al₂O₃和 TiO₂的含量(Gervasoni et al., 2017),这也可以解释为什么在高压条件下可以形成 石榴子石(6GPa)。同时,需要注意的是,硅不饱和 熔体同样会和地幔橄榄岩反应消耗斜方辉石,因此 交代形成的异剥橄榄岩或者地幔橄榄岩中出现的异

剥橄榄岩脉,既有可能是碳酸岩或富碳酸盐的硅质 熔体交代形成,也有可能是硅不饱和熔体交代形成 (Zinngrebe and Foley, 1995)。

碳酸岩熔体和富碳酸盐的熔体与地幔橄榄岩发 生反应,不同的一点就是富碳酸盐的硅质熔体和橄 榄岩反应还会形成 Fe-Ti 氧化物(Gervasoni et al., 2017)。这种现象在自然存在的岩石圈地幔捕掳体 中也很常见,常常被用来研究其所处岩石圈地幔的 氧逸度(fo₂)信息(Lorand and Gregoire, 2010)。 而克拉通岩石圈自太古宙形成之后一直经历着交代 作用,在交代作用过程中其氧化还原状态也会随之 改变(Canil et al., 1994; Foley, 2008, 2011; Frost and McCammon, 2008),这些含有 Fe-Ti 氧化物的 交代捕掳体也许可以用来指示长期以来深部岩石圈 地幔氧化还原状态的改变。同时也解释了为什么很 多碱性岩浆作用产生的岩浆具有较高的 Fe-Ti 含量。

3.2.4.3 流体交代作用

地幔交代中的流体可以整体划分为富水流体和 超临界流体(supercritical fluid)(Hermann et al., 2013; Zheng Jianping et al., 2019)。在俯冲体系 的岩浆岩系统中,富水流体在低压条件下(<3~ 4GPa)会与含水熔体独立存在,而在高压条件下会 与含水熔体混为一体形成超临界流体(Zheng Yongfei et al., 2011)。由于温度压力的增大,超临 界流体携带和迁移富集组分的能力比富水流体要高 好几个数量级(Schmidt and Poli, 2014; Zheng Jianping et al., 2019)。我们通常指的流体交代作 用一般出现在较浅的地幔源区,在较深的地幔区域 常常把流体/熔体作为一个整体来讨论交代作用的 机制。流体交代除了和地幔橄榄岩反应形成新的含 水硅酸盐矿物、碳酸盐矿物以及伴生的硫酸盐/硫化 物(显交代作用),还会造成地幔橄榄岩矿物相中不 相容元素和挥发分的富集,不会产生新的矿物相 (Scambelluri et al., 2006),即隐交代作用。通过自 然界存在的矿物(透辉石和石榴子石)包裹体的成分 研究显示交代流体不仅含有大量的水和 CO₂,还携 带大量的大离子亲石元素(LILE)和轻稀土 (LREE),因此在交代作用过程中会使地幔橄榄岩 富集 LILE 和 LREE,以及出现含水矿物相,如金云 母和角闪石(Wunder and Melzer, 2003; Tomlinson et al., 2009; Fumagalli et al., 2009).

以上介绍的几种交代作用往往是共同在岩石圈 地幔中作用的,正是大区域上多种交代作用的共同 参与,才会导致了岩石圈地幔的高度不均一性。同时,不仅是俯冲系统会产生这种交代富集的岩石圈地幔,软流圈和地幔柱也是在演化过程中释放富集流体和熔体与上覆岩石圈发生反应,产生不均一的交代富集地幔(Sen et al., 2009; Beccaluva et al., 2009; Chauvet et al., 2011; Howarth et al., 2014; Aviado et al., 2015; Saha et al., 2017),如岩石圈地幔中存在金云母橄榄岩、石榴子石-金云母橄榄岩、交代脉(异剥橄榄岩交代脉、金云母交代脉、辉石岩交代脉、角闪石交代脉)、MARID型交代地幔[云母(Mica)-角闪石(Amphibole)-金红石(Rutile)-钛铁矿(Ilmenite)-透辉石(Diopside)]等(O'Reilly and Griffin, 2013),可能就是俯冲系统和 深部地幔释放的流体/熔体共同作用的结果。

3.3 碱性煌斑岩

大部分碱性煌斑岩出露在离散型大陆边缘(如 大陆裂谷)和板内构造环境(如洋岛),通常和碱性玄 武质岩浆作用密切相关(Bernard-Griffiths et al., 1991; Gibson et al., 2006; Orejana et al., 2008; Batki et al., 2014; Tappe et al., 2016)。关于碱性 煌斑岩的岩石成因主要有两种模型,第一种模型和 之前所描述的钙碱性煌斑岩一样,可能来自一个交 代富集岩石圈地幔的低程度部分熔融,导致岩石圈 发生交代富集的组分可能来自于软流圈 (Aghazadeh et al., 2015; Hauser et al., 2010),也 可能来自于俯冲系统释放的流体或熔体 (Ngounouno et al., 2005; EI-ahmadi et al., 2011; Shukla et al., 2015)。第二种模型认为形成碱性煌 斑岩岩浆的源区是极度不均一的,而这种不均一的 源区组分包括经历交代富集的岩石圈地幔、深部熔 体与岩石圈地幔反应形成的富含挥发分的交代脉以 及上涌的深部地幔组分,其中,深部地幔组分又可以 分为地幔柱组分(Bernard-Griffiths et al., 1991; Gibson et al., 2006; Lundmark et al., 2011; Stoppa et al., 2014)和上涌软流圈组分(Orejana et al., 2008; Tappe et al., 2016; Raeisi et al., 2019)。碱性煌斑岩正是以上这些组分熔融产生的 熔体发生混合的结果。

正如之前所描述的碱性煌斑岩多与板内碱性岩 浆作用密切相关,所以常常用碱性岩浆作用的模型 来解释碱性煌斑岩的成因(Tappe et al., 2016 以及 所引的相关参考文献)。

来自大洋或大陆板内构造环境中的碱性岩石 (如碱性玄武岩、霞石岩和碧玄岩),它们通常具有比 大洋中脊玄武岩(MORB)含量高的不相容微量元素 组分(Gast, 1968),说明这些碱性岩石不可能来源 于一个均一亏损地幔橄榄岩源区。大洋和大陆板内 玄武岩变化的、富集的同位素组分说明它们来源于 一个不均一的、富集的地幔源区(Hofmann and White, 1982; Zindler and Hart, 1986; Chauvel et al., 1992; Morgan and Morgan, 1999; Ito and Mahoney, 2005; Tappe et al., 2016)。而产生碱 性岩浆源区的富集组分的性质和来源一直是一个具 有争议的话题(Lustrino and Wilson, 2007; Stracke, 2012; Zheng Yongfei, 2012)。关于源区 的富集有两种情况,一是俯冲循环洋壳士沉积物的 加入(Barling and Goldstein, 1990; Weaver, 1991; Chauvel et al., 1992),如海洋岩石圈在俯冲过程中 转化为辉石岩/榴辉岩,然后再循环到地幔对流中 (Hofmann and White, 1982; Hirschmann et al., 2003; Sobolev et al., 1996); 第二种不仅包括之前描 述的富集组分对岩石圈地幔的交代富集(Menzies and Murthy, 1980; Foley et al., 2001; Pilet et al., 2011; Dai Liqun et al., 2014),还包括俯冲作 用、拆沉作用或地幔柱作用过程释放的流体/熔体与 岩石圈地幔相互作用导致的富集(Halliday et al., 1995; McKenzie and O'Nions, 1995; Class and Goldstein, 1997; Niu Yaoling and O'Hara, 2003; Workman et al., 2004; Class and Le Roex, 2006)。最初的观点认为这两种富集方式是相互独 立的(Hofmann, 1997; Niu and O'Hara, 2003),但 不可否认的是这两种富集形式对解释产生板内岩浆 源区的地球化学和同位素组成都有着十分重要的作 用(Tappe et al., 2016)。产于大洋和大陆的板内 碱性岩浆不能通过"干"地幔橄榄岩(不含 H₂O 和 CO₂)部分熔融产生,即使这些岩浆的产生涉及到了 这些"干"地幔橄榄岩,那么也需要 CO₂ 的加入 (Eggler and Holloway, 1977; Wyllie, 1977; Dasgupta et al., 2010)。实验岩石学研究结果表明 含有 0.1%~0.25%CO2的地幔橄榄岩发生约 1% ~5%部分熔融就可以产生与黄长岩、霞石岩和碧玄 岩岩浆相似的熔体(Hirose, 1997; Dasgupta et al., 2007)。以上实验模拟的源区和熔融过程产生 的岩浆虽然大部分主量元素符合碱性岩浆的特征, 但是其 TiO₂和不相容微量元素含量达不到自然界 碱性岩浆的富集程度(Prytulak and Elliott, 2007)。 另外一种产生碱性岩浆的方式就是含挥发分的富角 闪石交代脉的部分熔融,这种方式产生的熔体可以 解释 OIB 和大陆碱性玄武岩的一些微量元素的同 位素特征(Pilet et al., 2011)。然而,这种"交代脉" 的成因模式需要这些"交代脉"的长时间演化解释这 些碱性岩浆的放射性同位素组成,即要求这种"交代 脉"必须在橄榄岩地幔中保留长达 1000Ma 之久 (Foley, 1992; Workman et al., 2004; Pilet et al., 2011; Tappe et al., 2008, 2013)。这种含水的交 代脉能否在对流的上地幔中如此长时间存留还存在 争议(Class and Goldstein, 1997; Frost, 2006)。 目前的研究认为,在相对较厚较冷的大陆岩石圈地 幔可能会长时间的保留这种交代脉(Menzies and Murthy, 1980; Hawkesworth et al., 1990b; Paslick et al., 1995)。同时, 岩石圈的厚度和热结 构对碱性岩浆的产生有很强的控制作用,碱性岩浆 也可能是交代富集的岩石圈地幔(含交代脉)、软流 圈(含相变的洋壳岩石圈,辉石岩)共同作用的结果 (Tappe et al., 2016).

4 煌斑岩研究过程中存在的一些问题

4.1 球粒结构和钠长石的成因

碱性煌斑岩中经常会出现的含碳酸盐和钠长石 的球粒,并且在基质中也会常常出现钠长石(Rock, 1991)。钠长石通常被认为是低温蚀变的产物 (Ferguson, 1960),虽然区别钠长石是岩浆阶段还 是蚀变阶段形成可以利用其有序度进行推断,但是 目前还没有关于碱性煌斑岩中钠长石有序度的报 道。根据之前研究的岩相学描述,煌斑岩中钠长石 的结晶程度往往较好,与其他岩浆期结晶的矿物呈 共结关系(Rock, 1991; Pe-Piper et al., 2018),这 说明钠长石形成于岩浆作用阶段,而不是岩浆期后 的蚀变阶段。然而根据 An-Ab 二元相图,我们可以 看出要结晶出钠长石需要体系极端富 Na(图 6),而 且在含H₂O条件下体系结晶出来的长石更富Ca (B'点)。同时,也没有高温高压实验证明在正常的 岩浆阶段可以结晶出钠长石,那么煌斑岩(主要为碱 性煌斑岩,Rock, 1991)基质和球粒中的钠长石是怎 么形成的呢?

作为典型的原生幔源岩石,除了后期蚀变形成 的杏仁构造,碱性煌斑岩中常常含有由碳酸盐或长 英质硅酸盐矿物组成的色率较低的球粒(Rock, 1991; Szabó et al., 1993; Azbej et al., 2006; Hauser et al., 2010; Batki et al., 2014)。煌斑岩 中的球粒结构通常被认为是:① 岩浆作用过程中两 种不混溶液相(硅酸盐-碳酸岩或硅酸盐-硅酸盐)发





The solid red line is the solid-liquid phase line of dry system (P=1 atm), and the blue dotted line is the solid-liquid phase line of water bearing system ($P_{H_2O}=5 \text{ MPa}$)

生分离结晶作用产生的(Philpotts, 1976; Eby, 1980; Fareeduddin et al., 2001; Vichi et al., 2005);② 富挥发分的岩浆演化晚期大部分基质结 晶之后,残余液相发生分离的再填充(Cooper, 1979; Foley, 1984; Azbej et al., 2006)。这两种 产生球粒的方式都可以理解为不混溶现象导致的, 只是这两种方式的不混溶现象发生的阶段不一样。 区分这两种情况形成的球粒结构有一个相对简单的 方法,那就是前者形成的球粒结构的边部矿物的组 成和地球化学成分(球粒富含碳酸盐、长英质矿物并 月贫镁铁质矿物)与岩石基质明显不同,而后者形成 的球粒结构具有分带现象,核部由方沸石或碳酸盐 组成,边部则具有与岩石基质相同的矿物组合,并且 不会有早期结晶矿物相出现(Foley, 1984),而且后 者形成的球粒结构相对于全岩具有较低的 TiO₂含 量(Philpotts 1976)。同时,早期岩浆不混溶产生的 液相会相对富 CaO(Kamenetsky and Kamenetsky, 2010),在冷却结晶过程中易形成富 An 的钙长石, 这与晚期液相形成富 Ab 的钠长石正好相反(Batki et al., 2014)。但是,这两个阶段形成的球粒有一 个共同点,那就是相比岩浆成因的碳酸岩,煌斑岩球 粒中的碳酸盐矿物具有较低的 REE 含量(Vichi et al., 2005).

由于大量含水镁铁质矿物相的出现以及缺乏长 石 斑 晶 说 明 煌 斑 岩 岩 浆 富 水 甚 至 是 水 饱 和 的 (Rock, 1991)。虽然这种富水的环境会促进不混溶 现象的发生(Leelanandam and Sreenivasan, 1986), 但是通常在高温高压条件下玄武质熔体-H₂O体系 是相互混溶的(Bureau et al., 1999),之后可能随着 岩浆演化过程和压力的降低,这种玄武质的岩浆会 水饱和进而发生流体-熔体的不混溶作用。不混溶 作用产生的不混溶流体由于岩浆的粘度比较大不容 易和硅酸盐熔体分离,往往以囊泡的形式或在早期 结晶的矿物缝隙中存在,产生球粒结构。这种球粒 也具有明显的分带现象,与晚期残余液相分离再填 充产生的球粒十分相似(Ballhaus et al., 2015)。 之前的讨论已经提到钠长石往往是低温蚀变阶段的 产物,这也就意味着钠长石是从流体相中析出的。 因此可以推测碱性煌斑岩碳酸盐球粒以及基质中的 钠长石可能是岩浆演化后期发生流体-熔体的不混 溶作用中流体相结晶析出的。这种流体不混溶模式 可以解释一些碱性煌斑岩同时出现球粒(以碳酸盐 矿物为主)和钠长石的现象,并且也可以解释为什么 这些碳酸球粒具有较低的 REE 含量。

4.2 钾质和钠质煌斑岩的源区

通常认为富集不相容元素和 LILE 并且亏损 HFSE 的岩石来源于一个经历过交代富集的地幔源 区,这种模式也被用来解释煌斑岩的成因(Foley et al., 1987, 2000; Tiepolo et al., 2000, 2001)。前 已述及,交代富集作用除了会使源区发生不相容元 素和 LILE 的富集,还可能会改变源区的矿物组合, 产生含水矿物相角闪石和黑云母(Gillis and Meyer, 2001)。目前,我们根据从岩相学、地球化学 和实验岩石学搜集到的数据可以将形成煌斑岩地幔 源区的矿物组合分为以下几种类型:① 金云母+ 单斜辉石 ± 橄榄石 (Esperança and Holloway, 1987; Wallace and Carmichael, 1989; Righter and Carmichael, 1996; Buhlmann et al., 2000; Elkins-Tanton and Grove, 2003; Foley et al., 2009; Parat et al., 2010; Funk and Robert, 2012); ② 云 母-单斜辉石岩 (Carlson and Nowell, 2001); ③ 富 金云母地幔橄榄岩 (Orejana et al., 2008); ④ MARID 型富集地幔(Tappe et al., 2008); ⑤ 富金 云母的类似 MARID 和碳酸盐/富金云母的混合源 区(Rosenthal et al., 2009)。以上源区中的金云母 相都用来解释煌斑岩富 K 的特征。然而,我们通过 搜集近几十年来发表的数据,显示煌斑岩可以分为 钠质煌斑岩和钾质煌斑岩。其中,划分钾质和钠质 的依据是:K₂O/(K₂O+Na₂O)>0.5 为钾质,K₂O/ (K₂O+Na₂O)<0.5 为钠质。这些富 Na 特征的煌 斑岩显然不能用源区的金云母相来解释,我们推测 是源区含有角闪石导致了煌斑岩富 Na 的特征。其 中,含角闪石的橄榄岩熔融产生的熔体具有较低的 Rb/Sr(<0.1)和较高的 Ba/Rb 比值(>20),而含金 云母的源区熔融产生的熔体具有较高的 Rb/Sr(> 0.1) 和较低的 Ba/Rb 比值(<20)(Latourrette et al., 1995; Furman and Graham, 1999)。我们根据 过去几十年发表的部分煌斑岩数据做 Rb/Sr-Ba/ Rb 相关图(图 7),忽略红色方框中的数据点(Rb/Sr <0.1 且 Ba/Rb<20),可以发现煌斑岩钠质和钾质 特征受源区金云母还是角闪石的影响比较明显。因 此我们认为富 K 煌斑岩的源区含水矿物相以金云 母为主,而富 Na 煌斑岩的源区含水矿物相以角闪 石为主。这种笼统的方法虽然解释了煌斑岩富 K 还是富 Na 特征受到了源区这两种含水矿物相的影 响和控制,但是源区具体的矿物相组合还要根据实 际情况进行讨论。



图 7 Rb/Sr vs. Ba/Rb 图解(据 Guo et al., 2004 修改) Fig. 7 Rb/Sr vs. Ba/Rb diagram of sodic and potassic lamprophyres (after Guo et al., 2004) K₂O/(Na₂O+K₂O)>0.5 为钾质煌斑岩, K₂O/(Na₂O+K₂O)<0.5 为钠质煌斑岩 Sodic lamprophyres: K₂O/(Na₂O+K₂O)>0.5, potassic lamprophyres: K₂O/(Na₂O+K₂O)<0.5

4.3 影响煌斑岩含水斑晶形成的因素

煌斑岩区别于其他岩石最显著的特征就是斑晶 由暗色矿物组成,并且斑晶主要是含水暗色矿物角 闪石和黑云母,有时甚至斑晶只有这两种矿物,不出 现橄榄石和辉石,正是以上的特征和较高的烧失量 (LOI)(原生碳酸盐矿物的存在也为煌斑岩的高 LOI做出了贡献),煌斑岩一直被认为是高含水量岩 石的代表(Rock, 1991)。实验研究表明岩浆中的水 会降低液相线温度并且能够显著地改变形成岩石中 的矿物相结晶顺序和比例(Hamada and Fujii, 2007)。较高的水含量也会抑制斜长石的结晶,使得 岩浆优先结晶出角闪石等矿物(Müntener et al., 2001),这就解释了为什么相比其他"干"基性岩石 (如玄武岩),煌斑岩中的斜长石不会以斑晶的形式 出现,只会出现在基质中。含水岩浆矿物结晶顺序 以及结晶过程中控制黑云母和角闪石结晶的主要因 素和条件是什么,一直是实验岩石学研究的重点 (Maaloe and Wyllie, 1975; Edgar and Arima, 1983; Elliott, 2001; Simakin et al., 2009; Kratzmann et al., 2014; Bucholz et al., 2012; Erdmann et al., 2014; Bucholz et al., 2014)。

对于黑云母,控制它早期结晶的主要控制因素 是熔体的主量元素组分,尤其是 K_2O 、Ti O_2 和 Al₂O₃。黑云母中K含量远远大于Na含量,因此高 K₂O和K₂O/Na₂O比值会有利于黑云母结晶条件 达到液相线并且使其优先结晶 (Edgar and Arima, 1983), TiO₂含量的增加和/或 Al₂O₃含量的降低会 提高黑云母的热稳定性(Righter and Carmichael 1996)。实验研究发现黑云母在熔体强水不饱和以 及低水含量的情况下也会发生结晶(Barton and Hamilton 1979; Esperança and Holloway 1987),相 反,控制角闪石结晶最主要的因素为熔体的水含量, 即使在水不饱和的情况下要使得角闪石发生结晶并 且稳定存在,那也需要熔体具有较高的水含量,大量 的实验研究表明,随着熔体水含量的增加,角闪石液 相线会向着角闪石优先结晶的方向移动(Holloway and Burnham, 1972; Müntener et al., 2001; Grove et al., 2003; Krawczynski et al., 2012). 例如一个花岗闪长质的熔体优先结晶出角闪石并且 保持稳定则需要相对较高的水含量 $(4\% H_2O, 0.2)$ GPa; 2.5% H₂O, 0.8 GPa), 而黑云母只需要较低 的水含量(< 0.5% H₂O)就能结晶并保持稳定 (Naney, 1983)。需要注意的是,由于黑云母是含水 矿物相,结晶时也需要一定量的水,所以在极低的水 压的条件下也是不能结晶的(Bucholz et al., 2014)。除了熔体的水含量和化学成分对角闪石和 云母的结晶起控制作用外,岩浆体系所处的温压条 件和氧逸度对这些矿物的结晶也会有影响。关于温 压和氧逸度对角闪石和云母影响的研究主要集中在 岛弧及俯冲岩浆体系的岩石,如流纹岩(Newman and Lowenstern, 2002), 安山岩 (Grove et al., 2002; Botcharnikov et al., 2006, Krawczynski et al.,2012),玄武岩(Zhang Bo et al.,2019),缺乏关 于煌斑岩岩浆中关于含水矿物结晶温压条件及氧逸 度的实验研究。我们可以根据角闪石和云母的化学 成分来反演它们结晶时的温压条件和氧逸度,这可 能对熔体为何能结晶出特定煌斑结构具有一定的指 示意义。

另外,虽然长期以来人们都认为岩浆岩中的含水的镁铁质矿物(角闪石和黑云母)是岩浆直接结晶 产物。然而,通过建模发现含水矿物不仅可以由岩 浆直接结晶形成,也可以通过含水熔体和现存无水 矿物相的反应产生,如辉石和含水熔体反应变为角 闪石或黑云母,这种现象在镁铁质岩石到中性岩石 会经常出现(Beard et al., 2004)。这一过程也可以 看作是脱水反应(Dehydration)的逆过程,即水合结 晶(Hydration Crystallization),其反应方程式为:含 水熔体+单斜辉石(Cpx)+斜方辉石(Opx)+Fe-Ti 氧化物±钙质斜长石=角闪石/黑云母+石英±钠 质斜长石(Beard et al., 2004)。

5 煌斑岩的成矿效应

煌斑岩不仅对研究地球的深部地幔过程,壳幔 相互作用以及板块构造有着十分重要的意义,而且 还与一些矿产有关,如金和金刚石矿床。

大量的研究显示钙碱性煌斑岩与金矿有着密切 的时空关系,世界范围内的许多金矿区均有煌斑岩 脉的出露,如加拿大苏必利尔省(Wyman and Kerrich 1989; Wyman et al., 2008; Mathieu et al., 2018), 澳大利亚伊尔岗克拉通(Barley and Groves 1990; Taylor et al., 1994; Vielreicher et al., 2015, 2016; Groves et al., 2018; Thébaud et al., 2018; Müller and Groves, 2018),中国云南等 (Wang Jianghai et al., 2001; Huang Zhilong et al., 2002; Lu Yongjun et al., 2013, 2015; Chen Yaohuang et al., 2014; Deng Jun et al., 2015; He Wenyan et al., 2016; Gan Ting and Huang Zhilong, 2017)。最初 Rock(1991)根据所测的数据 显示钙碱性煌斑岩的金含量比其他火成岩高一到两 个数量级,因此认为较高的含金量可能是钙碱性煌 斑岩的固有特征,而且形成交代富集地幔所需的含 挥发分和富集 LILE 流体也能够运输金(Rock et al., 1988a, 1988b)。然而之后的研究对煌斑岩和 金矿床之间的直接成因关系提出了质疑,尽管煌斑 岩的侵位时间和金矿床的形成时间非常一致,矿体 也的确沿着煌斑岩侵位的断层和剪切带分布,但是 附近没有矿化的钙碱性煌斑岩其本身并不富集金, 这说明富金并不是钙碱性煌斑岩固有特征,可能是 受到金矿相关的后期热液叠加导致(Taylor et al., 1994; Yeats et al., 1999)。Müller and Groves (2018)根据现有的地球化学和年代学数据统计,认 为仅仅处于相同的构造环境(地壳变形带)才会使全 球范围内造山型金矿床和煌斑岩之间呈现密切的空 间关系,而并没有直接的成因联系。

普遍认为原生经济型金刚石矿通常只会产出于 金伯利岩和钾镁煌斑岩这两种超镁铁质岩石类型中 (Helmstaedt and Gurney, 1995; Heaman etal., 2004)。其实在超镁铁质煌斑岩,碱性煌斑岩和钙碱 性煌斑岩中也发现了金刚石,部分甚至可以成为产 生经济效益的金刚石矿床,如加拿大拉布拉多和魁 北克地区(Tappe et al., 2008)、苏必利尔省 (Wyman et al., 2006, 2015), 阿比提比绿岩带 (Lefebvre et al., 2005)以及澳大利亚和南非等 (Rock, 1991)。目前关于金刚石的研究主要集中在 金刚石形成的温压条件和 C 的来源问题。普遍认 为金刚石形成于距地表 150~200km 即克拉通之下 的高压较低温地幔源区。地幔源区赋存金刚石的主 要有两种岩石,分别为地幔橄榄岩和榴辉岩,通过它 们的包裹体和C同位素的研究显示地幔橄榄岩中 的金刚石 C 来源于地幔, 而榴辉岩中的金刚石 C 来 源于俯冲板片中的地壳物质(Richardson et al., 1984, 1990; Haggerty, 1986; Richardson, 1986). 不管是金伯利岩、钾镁煌斑岩还是煌斑岩,大多数情 况下都认为金刚石的形成和它们并没有成因关系。 可能只是因为形成金伯利岩、钾镁煌斑岩和煌斑岩 岩浆的源区很深,恰好处于能够形成金刚石深度的 地幔源区,所以这些岩石常常作为寄主岩石将金刚 石沿岩浆通道快速携带至地表。而至于为什么只能 是这些岩石以及将金刚石携带至近地表的机制还不 是十分明确(Kirkley et al., 1991)。

如此之外,相对于大多数火成岩,煌斑岩由于富 集F、Cl、S、H₂O和CO₂,因此具有更强的溶解和运 输元素的能力。同时,这些挥发性成分也可能会在 煌斑岩岩浆上升的过程溢出岩浆,对围岩进行爆破。 随着这种过程的反复进行,围岩中的一些特定金属 元素(如Zr、Th、U、Mo、Sn等)会共同与挥发分再 次溶解到煌斑岩岩浆中,使得岩浆中的这些金属元 素进一步富集进而成矿(Rock, 1991)。尽管目前 还没有可靠的数据来证实或否认这一点,但是由于 目前对煌斑岩种类复杂并且成因机制还没有一个统 一的认识,因此关于煌斑岩的成矿效应还有待进一步去挖掘和探索。

6 总结和展望

本文系统地总结了煌斑岩的分类,岩石学特征 和岩石成因,同时提出了煌斑岩研究过程中存在的 一些科学问题。煌斑岩可以分为超镁铁质煌斑岩、 钙碱性煌斑岩和碱性煌斑岩。不同类型的煌斑岩产 出于不同的构造环境,可能也代表了不同的岩石成 因和形成过程,这对我们研究地幔的不均一性,富集 过程以及大规模构造运动具有十分重要的指示 意义。

(1) 煌斑岩类型的划分主要是根据岩石中的浅 色矿物,即碱性长石、斜长石和似长石。这些浅色矿 物只会出现在基质中,颗粒很小,因此在之前的研究 过程中会出现煌斑岩种类的误判。在今后的研究 中,对煌斑岩种类的限定需要谨慎,应当结合精确的 浅色矿物种类分析和所处的构造环境综合判定。正 确的种类判定会为之后煌斑岩的成因研究奠定良好 的基础。

(2)前人的研究重点主要集中在利用岩相学和 地球化学特征来揭示煌斑岩的成因,进而限制它们 所代表的动力学意义,而忽略了煌斑岩特殊的矿物 学以及结构构造所指示的意义。

(3)今后的工作中可以结合矿物、特殊结构构造 和实验岩石学对煌斑岩源区以及形成过程中的物理 化学条件进行限定,从而进一步对不同环境下地球 动力学过程改造地球深部的方式进行限制。

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Classification, characteristics and petrogenesis of lamprophyres: an overview

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Abstract

As some of the oldest recognized alkaline rocks, lamprophyres have been widely studied by geologists for hosting economic minerals such as gold, diamond. They are also significant in our understanding of the deep-mantle geodynamic processes. However, there are different understandings about the petrogenesis of lamprophyres. We summarized the research results of lamprophyres in recent decades as well as compiled a detailed overview of their classification, characteristics and petrogenesis in this paper. Lamprophyres can be divided into ultramafic lamprophyres, calc alkaline lamprophyres and alkaline lamprophyres according to the classification criteria from International Union of Geological Sciences (IUGS). Ultramafic lamprophyres are often the products of magmatism in an extension setting, which are associated with kimberlite and carbonatite. Calc-alkaline lamprophyres usually present in convergent and passive continental margin settings, and with several mechanisms of petrogenesis (e.g. differentiation of basic magma, magma mixing and low-degree partial melting of metasomatized mantle). Alkaline lamprophyres are usually closely related to alkaline basaltic magmatism from the divergent margins and intraplate tectonic settings. Regardless of the environmentand formation process of the rocks, these three types of lamprophyres are generally considered to be derived from a mantle source that has undergone metasomatic enrichment. Finally, we highlight some scientific problems in the study of lamprophyres, such as the influence of mineral phases (phlogopite and amphibole) in the enriched mantle source on the generation of sodic or potassic lamprophyric magma, the controlling factors in the formation of hydrous mineral phenocrysts, and the genesis of carbonate-rich "ocellar" texture and albite in the lamprophyres.

Key words: lamprophyres; classification; petrology; petrogenesis; mantle metasomatism