

部分熔融岩石流变学

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内容提要:本文系统地总结了静态与动态条件下部分熔融岩石中熔体的形态及其分布特征, 着重阐述部分熔融对橄榄岩和花岗岩流变学性质的影响。众所周知, 部分熔融不仅是造成地球的成分演化、形成层圈构造的重要过程, 而且对深部地壳和上地幔的物理性质(如, 电导率、滞弹性、剪切波速度和渗透性等)皆具有重要的影响。尤其是, 部分熔融岩石流变性的研究对于深刻理解地壳和岩石圈地幔之间的力学耦合、地幔对流, 板块构造运移、造山作用、地壳隧道流等地质过程必不可少。在过去特别是近三十年来, 该领域的实验和理论研究皆取得了长足的进步, 加深了人们对部分熔融岩石流变学性质的理解, 现已达成如下共识: 在静态条件下, 当橄榄岩中的熔体分数很小(< 约2%)时, 熔体主要局限于颗粒三连点或沿粒棱分布; 随着熔体体积分数的增加, 颗粒边界上的熔体膜也会逐渐增多。然而, 在共轴挤压和简单剪切变形条件下, 熔体大多数沿着与最大主应力呈 $15^\circ \sim 30^\circ$ 的小角度的伸展剪切带分布。只要熔体的体积分数不大(约5%), 绝大多数颗粒边界并没有被熔体润湿, 部分熔融只会导致中等程度的流变弱化效应。但是, 随着熔体含量的增加, 地幔岩流变强度的弱化效应会渐趋强烈, 变形最终导致固—熔体的彼此分离(岩浆萃取)。

关键词:部分熔融; 熔体形态与分布; 二面角; 流变学; 大应变剪切; 地球动力学

部分熔融及其熔体的萃取、聚集、迁移和演化是造成地球成分演化形成层圈构造(Defant and Drummond, 1990; Lay et al., 2004; Anderson, 2007)的重要地质过程, 而且对深部地壳和上地幔的物理性质, 如电导率(Shankland and Waff, 1977; Shankland et al., 1981; Schmeling, 1986; Watanabe and Kurita, 1993)、滞弹性(Schmeling, 1985; Faul, 2004; Jackson et al., 2004)、弹性波速度(Stocker and Gordon, 1975; Sato and Sacks, 1989; Muellor and Raab, 1997; Hammond and Humphreys, 2000; Yang Xiaosong et al., 2003)、渗透性(von Bargaen and Waff, 1986; Faul et al., 1994)和流变性(Hirth and Kohlstedt, 1995a, b; Ji Shaocheng, 2008)等皆具有非常重要的影响。洋中脊之下的上地幔的部分熔融的程度直接影响海底扩张和洋壳形成的速率, 岛弧火山活动是俯冲板块之上地幔楔内部分熔融的

产物, 是造山带内大量花岗岩的形成与地壳增厚、深部热能增加导致岩石部分熔融的结果(杨晓松和金振民, 1999; 马麦宁等, 2002)。天然的部分熔融物质作为一种多相材料(Ji Shaocheng and Xia Bin, 2002; 周永胜等, 2003; 周平等, 2006), 其中的熔体相不仅刚度为零而且体积模量一般都较低(Karato, 2008), 以致固—液相之间巨大的力学反差使得整个体系的流变学性质变化极大(Ji Shaocheng, 2008)。再说, 无论在共轴挤压还是在简单剪切变形的条件下, 熔体含量及其分布对部分熔融岩石的力学性质的影响是巨大的。因此, 有关部分熔融岩石流变学性质的研究成果对于深刻理解地球各圈层的变形、地幔对流、板块构造运移、造山带形成和地壳隧道流等地质过程至关重要。

在过去特别是近三十年来, 部分熔融岩石流变学领域的实验和理论研究皆取得了长足的进步, 加

注: 本文为中国科学院知识创新工程重要方向项目(编号 KZCX2-YW-Q08-03-4), 国家自然科学基金项目(编号 40772029), 中国科学院“百人计划”项目(编号 KZCX-0733221A07)、青藏高原东部和东南部岩石圈流变学及地球物理性质研究项目(编号 1212011121274)、国土资源部深部探测技术与实验研究专项课题(编号 SinoProbe-07)、中国科学院知识创新工程领域前沿项目(编号 GIGCX-09-02)的成果。

收稿日期: 2011-05-24; 改回日期: 2011-09-06; 责任编辑: 章雨旭。

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深了人们对部分熔融橄榄岩和花岗岩流变学性质的理解,现已达成许多共识。本文将系统地总结静态与动态条件下部分熔融岩石中熔体的形态及其分布(拓扑结构)的特征,着重阐述部分熔融对橄榄岩和花岗岩流变学性质的影响,为解决当前地质界非常关心的有关地球动力学的疑难问题提供新的思路。

1 静态条件下的熔体形态与分布

部分熔融岩石的物理和流变学性质主要取决于熔体润湿颗粒边界的程度 (Stocker and Gordon, 1975; 侯渭等, 2004; 周平等, 2006), 其中二面角 (Dihedral angle, θ) 或润湿角 (Wetting angle) 是一

个至关重要的参数 (Cmiral et al., 1998; Mibe et al., 1998)。根据熔体分布理论, 岩石中主要有三种熔体分布方式: 当 $\theta > 60^\circ$ 时, 熔体在颗粒拐角处呈孤立的熔体包 (melt pocket) (图 1a); 当 $0^\circ < \theta \leq 60^\circ$ 时, 熔体从三连点出发并沿着粒棱分布, 构成熔体管道 (图 1b); $\theta = 0^\circ$, 熔体呈薄层或薄膜形式沿颗粒边界分布, 形成三维的熔体网络 (图 1c)。即使在二维的切片或薄片上确实观察到沿颗粒边界分布的熔体薄层或薄膜 (Jin Zhenming et al., 1994), 也不能据此断言熔体布满了所有的晶面 (图 1d), 因为在晶体与晶体的接触点或接触面上依

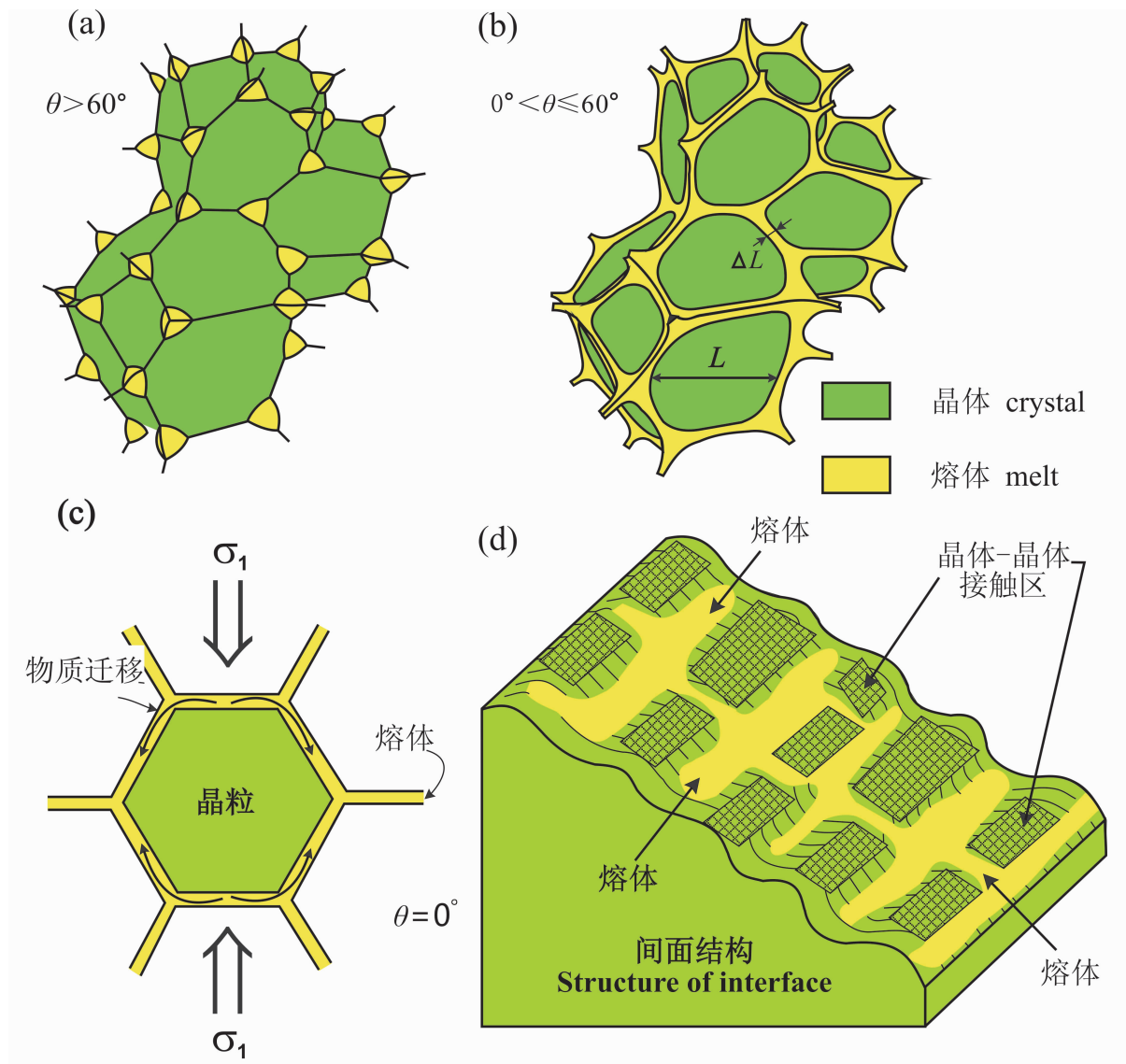


图1 部分熔融岩石中熔体的二面角类型:(a)二面角大于或等于 60° ;(b)二面角小于 60° ;(c,d)二面角接近 0° 。注意在晶体—晶体的接触面上没有熔体(d)

Fig. 1 Main types of partially molten rocks with the dihedral angle larger than or equal to 60° (a), less than 60° (b) and close to 0° (c and d). No melt occurs at the crystal—crystal interfaces (d)

然没有熔体。熔体的联通性还直接影响部分熔融岩石中熔体的迁移、不相容元素在固—液两相之间的分配 (Hart, 1993; Zhu Wenlu et al., 2011)。

在各向同性的静水压力条件下,部分熔融岩石中熔体的二面角是受固—液和固—固界面自由能的相对大小控制的 (图 2a), 分别记着 γ_{sl} 和 γ_{ss} (Kohlstedt, 2002; Karato, 2008)。熔体的二面角 θ 由下式给出:

$$\cos \frac{\theta}{2} = \frac{\gamma_{ss}}{2\gamma_{sl}}$$

(Watson et al., 1990; Kohlstedt, 1992, 2002; Karato, 2008):

(1) 当 $\gamma_{ss} \geq 2\gamma_{sl}$, $\theta = 0^\circ$ 。

(2) 当 $\gamma_{sl} < \gamma_{ss} < 2\gamma_{sl}$, 即 $0^\circ < \theta \leq 60^\circ$ 时, 熔体中的流体压力比外界的要小。

(3) 当 $\gamma_{ss} < \gamma_{sl}$, $\theta > 60^\circ$ 时, 只要熔体的体积分数不大, 熔体呈孤立的熔体包出现, 熔体包内的流体压力会大于外界的压力 (Cooper, 1990; Brenan and Rose, 2002; Kohlstedt, 2002; Karato, 2008; Yoshino et al., 2009)。

对于大多数的硅酸盐体系, 二面角一般为 30°

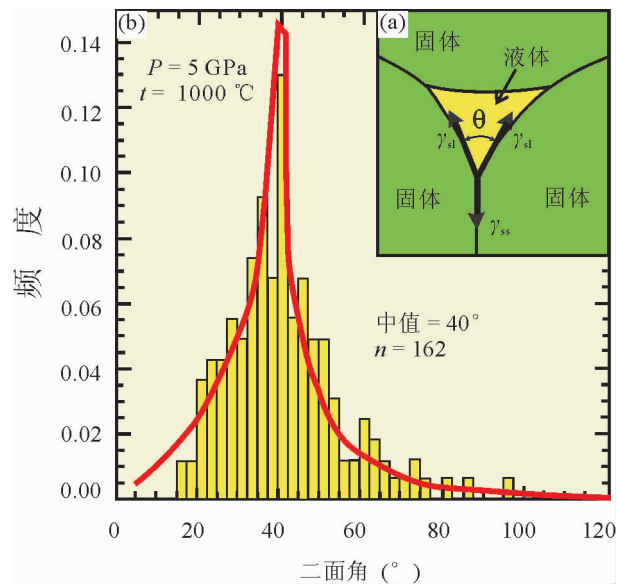


图 2 (a) 二面角与界面能的关系; (b) 5 GPa and 1000 °C 时橄榄石多晶集合体中含水 (5%) 流体的二面角频度分布直方图, 中值为 40° 。据 Mibe 等 (1998)

Fig. 2 (a) The relation between dihedral angle and interfacial energies; (b) histogram for the frequency distribution of dihedral angles at aqueous fluid in polycrystalline olivine aggregates at 5 GPa and 1000 °C. After Mibe et al. (1998)

$\sim 50^\circ$ (Waff and Bulau, 1979, 1982; Waff, 1980; Cooper and Kohlstedt, 1982; Toramaru and Fujii, 1986)。图 2b 所示的是在温度 1000 °C 和压力 5 GPa 的条件下热压合成的橄榄石多晶集合体中含水 (5.0%) 熔体的二面角分布图, 中值为 40° (Mibe et al., 1998)。石英—斜长石体系 (水含量小于 0.2%) 部分熔融的二面角为 $55^\circ \sim 60^\circ$ (Jurewicz and Watson, 1985)。角闪岩部分熔融的二面角大于 60° (Wolf and Wyllie, 1991); 低压 ($P < 2$ GPa) 时斜长岩部分熔融的二面角也会大于 60° (Watson and Brenan, 1987; Laporte and Watson, 1991)。

熔体的二面角取决于温度、压力条件 (Takei and Shimizu, 2003; Yoshino et al., 2002, 2007, 2009)、晶面方位 (Waff and Faul, 1992; Schafer and Foley, 2002), 以及与固体共存的熔体的化学成分 (Wanamaker and Kohlstedt, 1991; Takei and Shimizu, 2003)。例如, Mibe 等 (1999) 与 Yoshino 等 (2007) 发现, 在橄榄石—水体系中, 二面角随着静水压力的升高而系统地降低 (图 3), 其中随着静水压力从 1 GPa 增加到 5 GPa 二面角从 46° 减小到 18° (Yoshino et al., 2007), 这一变化趋势在较低温度 (约 1000 °C) 下的相同体系的实验 (Mibe et al., 1998) 中也有见到。此外, 从图 3 中还可以发现, 在橄榄岩中的玄武岩熔浆的二面角随着温度升高而减小 (Yoshino et al., 2009)。岩石中已有的矿物晶格优选定向 (LPO) 或晶形优选定向 (SPO) 也能显著地改变熔体网络的几何形态 (Cooper and Kohlstedt, 1984; Waff and Faul, 1992; Faul et al., 1994; Zimmerman and Kohlstedt, 2004), 这是因为熔体往往优先润湿那些低指数的晶面 (Cooper and Kohlstedt, 1982; Waff and Faul, 1992; Faul et al., 1994), 例如橄榄石的 (010) 面。如果熔体和固体的化学成分相近, 其二面角必然就小 (Urai, 1983; Urai et al., 1986; Mibe et al., 1998; Takei and Shimizu, 2003; Yoshino et al., 2002, 2007), 熔体就会润湿大多数的颗粒边界。此外, 在含变价元素的体系中, 熔体的二面角还会随着氧逸度的增加而有所改变 (Minarik et al., 1996; Gaetani and Grove, 1999)。

理论上说, 上述三种熔体形态与分布的类型在天然和人工合成材料中均能出现 (Cooper, 1990), 但是在以硅酸盐矿物为主的地壳和地幔岩石中, 熔体的分布方式主要为第 (2) 种, 例如在橄榄岩中玄武岩熔体的二面角一般为 $30^\circ \sim 50^\circ$ (Waff and

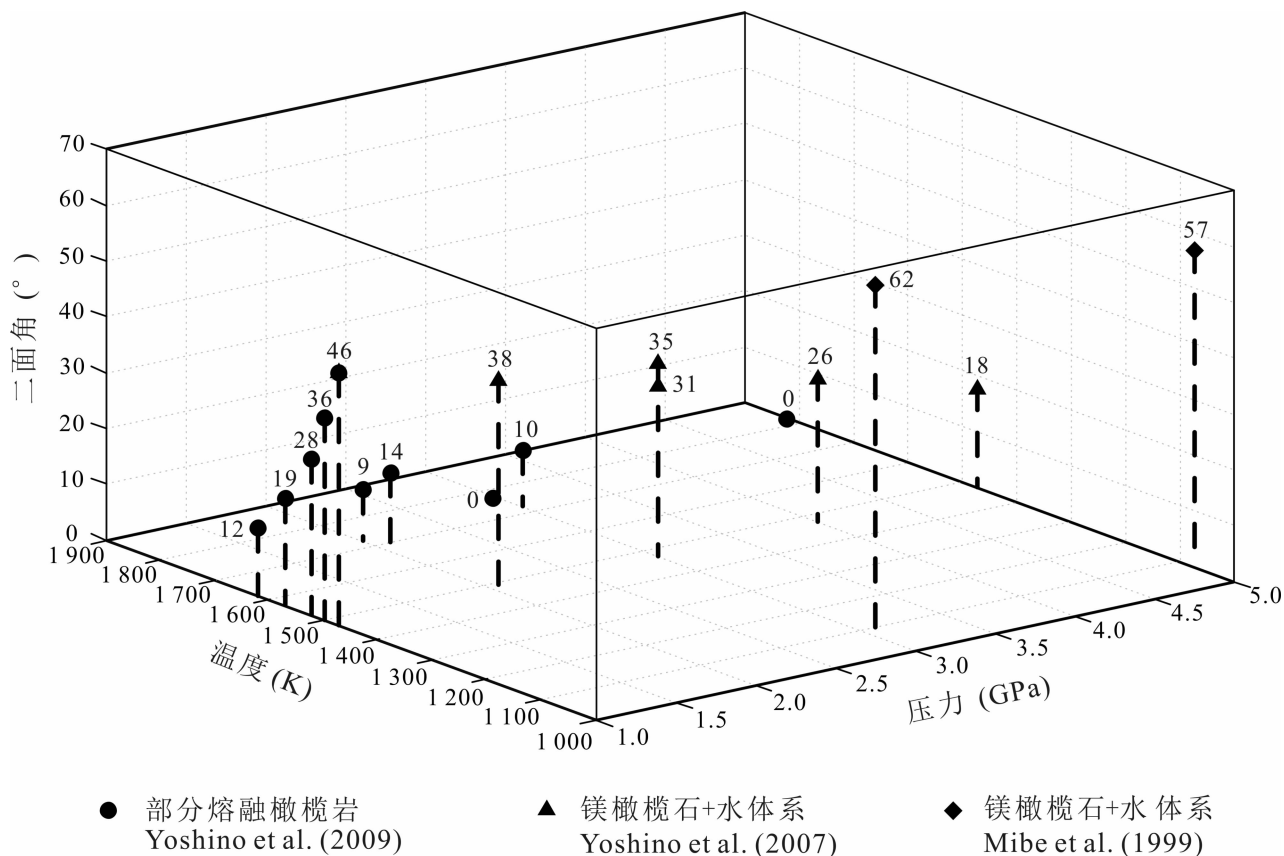


图3 部分熔融橄榄岩 (Yoshino et al., 2009) 和镁橄榄石—水体系 (Mibe et al., 1999 与 Yoshino et al., 2007) 中二面角随温度和压力的变化

Fig. 3 Dihedral angle variations in partially molten peridotites (Yoshino et al., 2009) and in the forsterite—H₂O systems (Mibe et al., 1999 and Yoshino et al., 2007) as functions of pressure and temperature

Bulau, 1979); 在地壳的中酸性或基性岩石中熔体的二面角一般为 $10^{\circ} \sim 30^{\circ}$ (Laporte et al., 1997; Dimanov et al., 2000), 因而可以认为绝大多数岩石的颗粒边界是没有熔体的。例如, Rybacki and Dresen (2000) 在研究由 Paterson 高温高压流变仪 (邵同宾等, 2011) 高温静压 (HIP: Hot isostatic pressing) 合成的钙长石多晶集合体时发现, 湿的样品中的熔体体积分数 $< 0.7\%$, 呈孤立的熔体包, 透射电镜 (TEM) 观察也没有发现颗粒边界上有熔体薄膜。此外, 对钙长石—透辉石多晶集合体 (Dimanov et al., 2003; Dimanov and Dresen, 2005) 和拉长石多晶集合体 (Dimanov et al., 1998) 的 HIP 实验也得出类似的结论。在这些作者所研究的热压合成的拉长石集合体中, 尽管熔体体积分数高达 12% , 熔体仍呈孤立的四边体分布, 颗粒边界依然没有发现熔体薄膜 (Dimanov et al., 1998), 这与后来在相同温度 (约 950°C) 条件下热压得到的拉长

石多晶集合体中观察到的熔体形态与分布类似 (Dimanov et al., 2000)。然而, Dimanov 等 (2000) 还发现在约 1150°C 热压得到的拉长石多晶集合体中, 除了有 2.0% 体积的熔体呈孤立的三角形分布外, 它还包含极少量 (约 0.1% 体积) 的熔体薄膜, 其厚度仅为 $5 \sim 50 \text{ nm}$ 。这些薄膜熔体的成分有别于纯的拉长石, 极其富含 SiO_2 (高达 80%)。因为热压的初始材料是拉长石玻璃, 薄膜熔体其实是尚未来得及完全结晶的初始玻璃, 这样非平衡的热压结构在自然界真实的地质材料中极可能是不存在的。

最近, Zhu Wenlu 等 (2011) 研究了静压条件下部分熔融橄榄岩中熔体的三维分布规律, 发现熔体主要沿着粒棱 (Grain edges) 分布, 他们将此称之为熔体管道 (Melt channels)。熔体管道进一步构成彼此连接贯通的熔体网络 (Melt network)。他们还发现, 当熔体体积分数较低 (2% 和 5%) 时, 只有小部分的颗粒边界被熔体润湿, 而绝大部分熔体从三

连点出发沿着粒棱形成熔体网络。只有当熔体体积分数增加到10%和20%时,熔体才会沿着一部分颗粒边界以薄膜的形式出现。这就是说,熔体的几何形态与分布随岩石中熔体的体积分数变化而变化,类似的情况在部分熔融的辉长岩中亦有报道(Zhou Yongsheng et al., 2011)。

2 动态条件下的熔体形态与分布

地壳和上地幔中变形岩石的部分熔融自始至终有差应力的参与,虽然差应力因地因时而异。应力和应变能有效地改变部分熔融岩石中熔体的形态和分布(Kohlstedt and Zimmerman, 1996; Zimmerman et al., 1999; Dimanov et al., 2000; Kohlstedt, 2002; Holtzman et al., 2003a; Groebner and Kohlstedt, 2006; Holtzman and Kohlstedt, 2007; Rybacki et al., 2008; King et al., 2010; Kohlstedt et al., 2010)。换句话说,部分熔融岩石中熔体的拓扑结构既可受应力也可受应变控制。

2.1 应力的影响

差应力不仅可以造成化学势的各向异性(Cooper and Kohlstedt, 1984; von Bargen and Waff, 1986; Hirth and Kohlstedt, 1995a),而且在脆性场和半脆性场内还能形成微破裂的优选定向排列(Drury and Roermund, 1989)。研究发现,在部分熔融岩石的实验变形过程中,如果有效压力较低或差应力较高都会在试样中产生张裂隙。例如,van der Molen and Paterson (1979)和Dell'Angelo and Tullis (1988)在低围压下对花岗岩的共轴挤压、半脆性变形的实验中观察到,熔体充填的颗粒边界或微破裂基本平行于最大主应力 σ_1 的方向,或垂直于最小主应力 σ_3 的方向。类似的情况也出现在共轴挤压变形的部分熔融尖晶石二辉橄榄岩中(Ave' Lallemant and Carter, 1970; Kohlstedt and Zimmerman, 1996)。基于上述事实,再据应力对界面化学势影响的分析,von Bargen and Waff (1986)估计只要上地幔中的差应力超过几个MPa,就能导致玄武岩熔体从母岩内萃取并分离出来,汇进平行于 σ_1 方向的裂隙中。但是,在简单剪切的塑性变形过程中,部分熔融岩石(如橄榄岩)中的熔体包会被压扁拉长,形成强烈的熔体形态优选方位(MPO: Melt preferred orientation),熔体包的长轴或最大压扁面的方向一般总是与最大主应力或主剪切面呈 $15^\circ \sim 30^\circ$ 的小角度相交(图4a和4b; Bussod and Christie, 1991; Kohlstedt and Zimmerman, 1996; Daines and

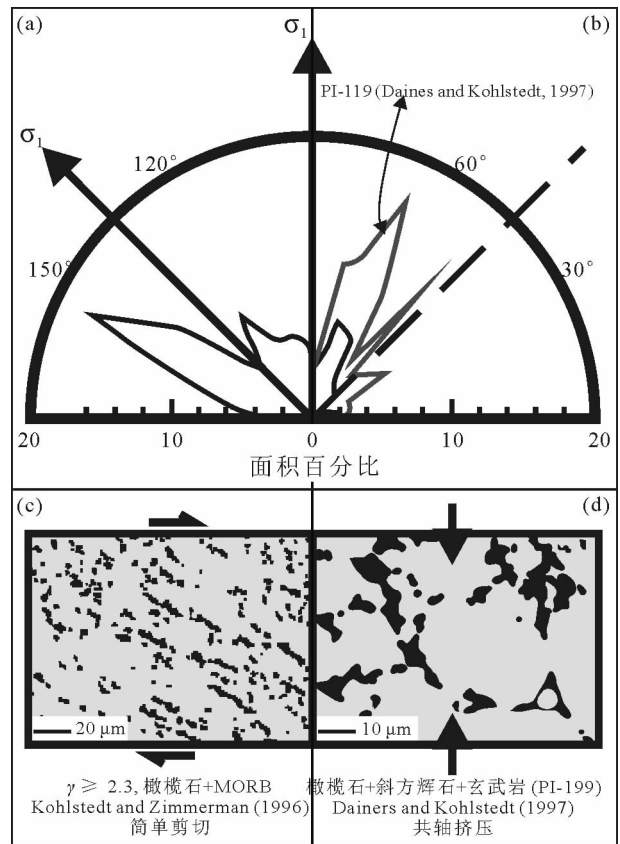


图4 干橄榄石 + MORB (a 和 c) 与橄榄石 + 大约4%的斜方辉石 + 玄武岩 (b 和 d) 分别在简单剪切 (Kohlstedt and Zimmerman, 1996) 和轴向挤压 (Daines and Kohlstedt, 1997) 条件下的变形显微构造。(a, b) 熔体包与最大主应力 σ_1 方向之间几何关系的玫瑰图;(c, d) 熔体(黑色)分布图。a 和 c 指示同一样品, b 和 d 分别指样品 PI-119 和 PI-199

Fig. 4 Microstructures of the samples for olivine plus MORB (a) and (b) and olivine with about 4% orthopyroxene plus basalt (b) and (d) deformed in simple shear (Kohlstedt and Zimmerman, 1996) and axial compression (Daines and Kohlstedt, 1997), respectively. (a) and (b) Rose diagrams of melt pockets vs. the orientation of the maximum principal stress under the two types of deformation. (c) and (d) binary images of the deformed samples. (a) and (c) illustrate the same sample while (b) and (d) display samples of PI-119 and PI-199, respectively

Kohlstedt, 1997; Zimmerman et al., 1999; Zimmerman and Kohlstedt, 2004),其剪切旋向也与试样整体的剪切旋向相同,说明熔体在非共轴或简单剪切变形过程中要逐渐地向伸展剪切带(ESB: Extensional shear band)内汇聚。类似的现象在自然

界高温剪切变形的混合岩中十分常见(图5),浅色花岗岩成分的部分熔体从面理分离出来涌进伸展剪切带(Brown, 1994)。

Urai (1983) 与 Urai 等 (1986) 发现,在岩盐—盐水体系的塑性变形过程中,流体湿润几乎所有的颗粒边界,这样的显微构造加快了颗粒边界迁移即动态重结晶作用。但当差应力撤除之后,流体又会重返静态时的分布样式。Jin Zhenming 等 (1994) 报道,在中等围压(约 1.5 GPa)的变形条件下,含低熔体分数的部分熔融橄榄岩具有与上述岩盐—盐水体系类似的显微构造,他们认为差应力导致玄武岩熔体润湿几乎所有的橄榄石的颗粒边界,因而导致强烈的弱化,但是并没有查明熔体润湿颗粒边界以及弱化的具体的物理化学机制,也没有观察到变形后长时间(约 40 h)的静态退火对熔体分布的显著影响。值得注意的是,在与 Jin Zhenming 等

(1994) 的实验相同的温度、位移速率和熔体分数、但围压较低条件下对与之组分类似的尖晶石二辉橄榄岩的实验观察仅见到极少量的沿颗粒边界分布的熔体薄膜(Kohlstedt and Zimmerman, 1996; Daines and Kohlstedt, 1997; Zimmerman and Kohlstedt, 2004)。迄今为止, Jin Zhenming 等 (1994) 的实验结果与 Kohlstedt 团队(Kohlstedt and Zimmerman, 1996; Daines and Kohlstedt, 1997; Kohlstedt, 2002; Zimmerman and Kohlstedt, 2004) 的实验结论之间差异的原因不详,值得探究。此外,美国的洋中脊 MELT 地震研究计划(The MELT Seismic Team, Science, 1998) 的集成性成果亦证明洋中脊以及两侧上地幔中的部分熔融是呈不均匀分布的,故导致测量到的地震波各向异性和剪切波分裂(Wolfe and Solomon, 1998)。

2.2 应变的影响

应变也是影响熔体的几何和重新分布的重要因素。一方面,随着应变量的增加,熔体包的长短轴会分别向 σ_1 和 σ_3 的方向(共轴应变)或剪切方向和剪切面的法线方向(简单剪切)旋转;另一方面,随着应变量的增加,位错滑移形成的矿物的晶格优选定向(LPO)和晶形优选定向(SPO, 即面理和线理)更加强烈。例如,在橄榄石作(010)[100]滑移时,(010)晶面平行或近乎平行于面理,而[100]晶格方向平行或近乎平行于拉张线理。受晶面自由能的控制,熔体总是优先润湿橄榄石的(010)晶面,故玄武岩熔体也就不可避免地形成了平行于面理的优选定向,Waff and Faul (1992)、Kohlstedt and Zimmerman (1996) 和 Daines and Kohlstedt (1997) 等特别强调玄武岩熔体在部分熔融地幔岩中分布的各向异性,认为 MPO 进而导致整体岩石流变强度的各向异性以及地震波速的各向异性。特别需要指出的是,能够影响矿物组构样式的因素有很多,例如差应力、应变方式(共轴挤压、简单剪切)及其大小、变形条件(温度、压力、剪应力大小、水含量)和变形机制等。实验发现, MPO 在简单剪切的试样(如, Kohlstedt and Zimmerman, 1996; Zimmerman et al., 1999) 中总是比在共轴挤压变形的试样(Daines and Kohlstedt, 1997; Hustoft et al., 2007) 中更为显著(比较图4中的c与d),其原因就是两者的应变不同,一般简单剪切($\gamma \geq 1$)比起共轴挤压($\epsilon \leq 0.2$)更容易获得较大的应变。

图6表示自然界熔体萃取的一种常见的、简单

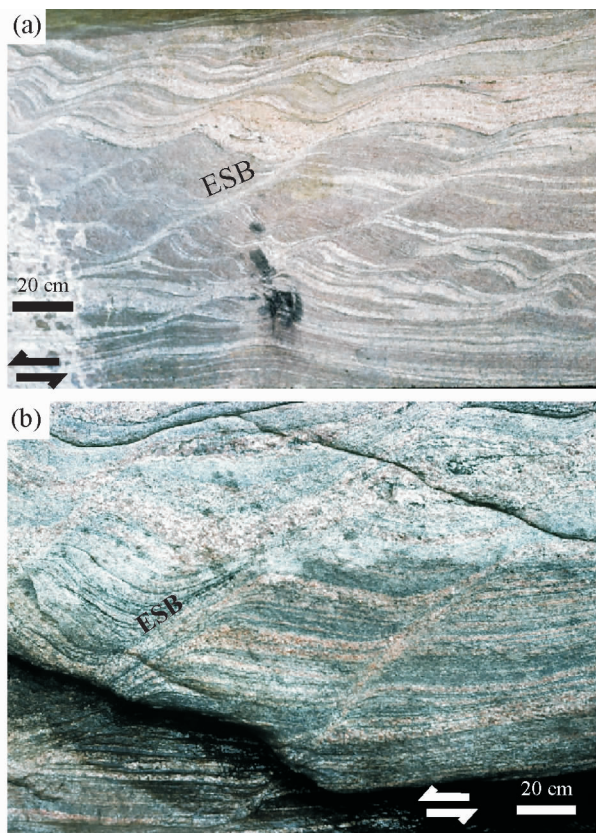


图5 来自加拿大 Grenville 构造省典型下地壳部分熔融岩石的信息。在(a)、(b)所示左旋剪切变形过程中,混合岩中的部分熔体(浅色花岗岩)从面理分离出来进入伸展剪切带(ESB)中

Fig. 5 Melt, now represented by leucosome, interpreted to have migrated along the foliation and into extensional shear bands (ESB) within migmatites (Grenville Province, Canada) deformed by sinistral shear (a) and (b)

可行的模式 (Richter and Mckenzie, 1984), 部分熔融岩石可以看作是一种被流体充填的多孔材料, 在塑性变形过程中固体介质发生压实, 挤出孔隙中的熔体, 由于熔体的密度较低故要上浮集中、汇聚成带。在实验中亦发现, 当应变量积累到一定程度之后, 部分熔融岩石中的熔体就会与熔体亏损的残余体分离并汇聚成带 (Holtzman et al., 2003a, b; Holtzman and Kohlstedt, 2007; Rybacki et al., 2008; King et al., 2010; Kohlstedt et al., 2010), 相邻两条富熔体带之间是熔体亏损带。理论模拟 (Stevenson, 1989) 和野外地质 (Kelemen and Dick, 1995) 也观察到类似的现象。在变形试样中, 富熔体带和熔体亏损带的相对厚度与熔体体积分数、压实长度 (h_c) 甚至颗粒粒度的大小皆具有一定的定

量关系 (Holtzman et al., 2003a; Holtzman and Kohlstedt, 2007; Kohlstedt and Holtzman, 2009; Kohlstedt et al., 2010)。富熔体带的宽度 (h_m) 和间距 (h_s) 随 h_c 和熔体体积分数 (φ) 的增加而增加。在粒度为约 $10 \mu\text{m}$ 的橄榄岩试样中, 富熔体带宽约 $50 \mu\text{m}$, 熔体亏损带宽 $100 \sim 200 \mu\text{m}$ 。在地下壳和上地幔, 熔体带和熔体亏损带的厚度要比实验室内标本中的大得多。

3 部分熔融岩石的变形

岩石部分熔融和变形是相互作用的, 一方面熔体受到差应力和变形的影响会发生重新分布 (Urai, 1983), 另一方面熔体的分布或润湿行为又对部分熔融岩石的变形和流动强度产生重要的影响 (Pharr and Ashby, 1983)。具体的说, 部分熔融岩石的流变强度主要取决于熔体润湿颗粒边界的程度 (Stocker and Gordon, 1975), 即熔体二面角的大小。对于上地幔橄榄岩, 玄武岩熔体的二面角一般为 $30^\circ \sim 50^\circ$, 表明熔体主要出现在颗粒三连点及粒棱, 并没有润湿全部的颗粒边界 (Waff, 1980; Toramaru and Fujii, 1986; Zhu Wenlu et al., 2011)。这方面研究工作做的最多的是美国明尼苏达大学 David L. Kohlstedt 教授领导的团队, 他们系统地研究过玄武岩熔体在不同实验条件下对部分熔融橄榄岩塑性变形和流变强度的影响 (Cooper and Kohlstedt, 1984; Hirth and Kohlstedt, 1995a, b; Kohlstedt and Zimmerman, 1996; Zimmerman, 1999; Mei Shenghua et al., 2002; Scott and Kohlstedt, 2006)。他们发现, 只要熔体的体积分数不大 ($< 8\%$), 绝大多数颗粒边界就不可能被熔体全部润湿 (图 7), 部分熔融只会导致中等程度的流变弱化。具体地说, 每增加 1% 体积的熔体, 岩石的流动强度降低约 $10\% \sim 40\%$ 。此外, 在相同熔体体积分数时, 颗粒径越小其界面愈容易被熔体润湿。例如, 试样 PI-73 中的熔体体积分数比试样 PI-19 要少约 1% , 但是由于前者的粒径 (约 $10 \mu\text{m}$) 比后者 (约 $12 \mu\text{m}$) 要小, 因此前者的颗粒边界被熔体润湿的比例要比后者略高一些 (图 7, Hirth and Kohlstedt, 1995a)。

Kohlstedt (2002) 重申了 Cooper and Kohlstedt (1986) 和 Cooper 等 (1989) 的熔体平衡的理论模式 (即, CK 模型), 该模式认为, 在扩散蠕变域, 部分熔融之所以能促进变形是因为物质在熔体中的迁移扩散速率远比在固体中要大。而在位错蠕变域, 熔

熔体从变形基质中分离

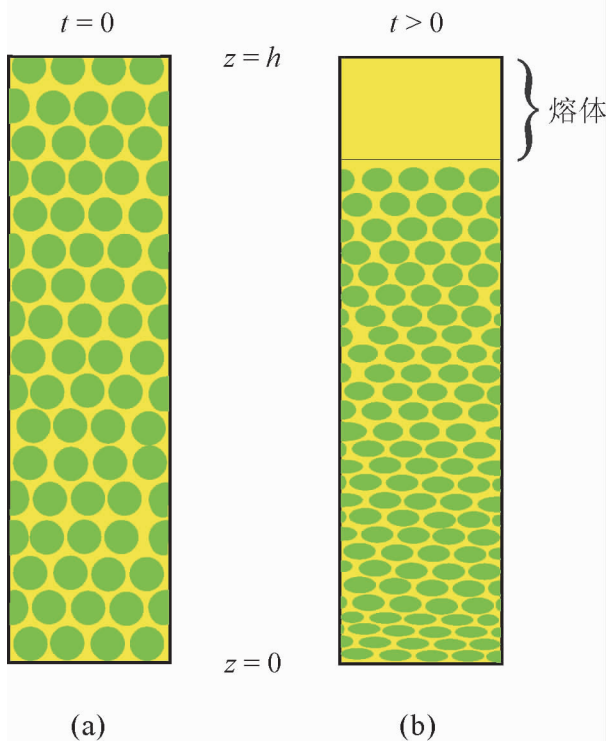


图 6 熔体萃取机制的示意图

Fig. 6 Schematic illustration of a dynamic model for melt segregation from a deformable, porous matrix (a)、(b) 分别表示熔体萃取前后的状态。高密度的固体发生塑性变形和压实、低密度的熔体遭挤出、上浮、汇聚起来 (改自 Richter and Mckenzie, 1984)

The (a) and (b) panels show, respectively, the states before and after the segregation of less denser melt, which is caused by progressive ductile deformation and compaction of the solid matrix. A melt layer forms at the top of the system. Modified from Richter and Mckenzie (1984)

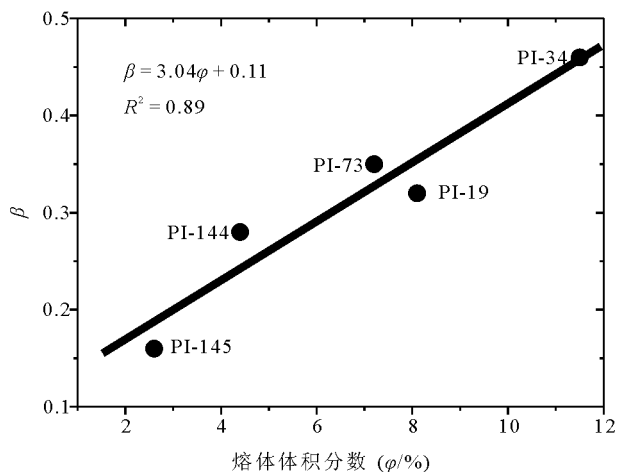


图7 颗粒边界被熔体完全润湿的比例系数(β)随熔体体积分数(ϕ)的变化(建图数据取自Hirth and Kohlstedt, 1995a)

Fig. 7 The fraction of two-grain boundaries completely wetted by melt (β) as a function of melt volume fraction (ϕ) (Data from Hirth and Kohlstedt, 1995a)

体依靠造成局部应力集中来促进塑性变形。无论上述哪一种变形机制,低程度的

部分熔融对以橄榄石为主要造岩矿物的地幔岩的流变强度的影响都不会太大,因为此时熔体还主要分布在颗粒三连点及粒棱,并没有润湿全部的颗粒边界(Karato, 2010),上述结论与Jin Zhenming等(1994)的截然不同,但得到了学界的普遍认同。

Yoshino等(2009)在围压 ≥ 3.0 GPa、温度 ≥ 1750 K的静压实验中发现玄武岩熔体润湿较大部分橄榄岩的颗粒边界,二面角小于 5° 。因为是静压实验,他们无从了解熔体润湿颗粒边界对整体岩石流变强度的影响。他们的实验不同于Jin Zhenming等(1994)在围压等于1.5 GPa、温度1475~1500 K的共轴挤压的差应力变形的实验结果,但是却意外地报道了相同的熔体分布结构。Yoshino等

(2009)的结论是,温度特别是压力的增加是造成熔体二面角减小的关键因素;而Jin Zhenming等(1994)的结论却是,变形使得熔体在岩石中作均匀分布,熔体二面角变为零是变形的结果,而不受温压条件的制约。Jin Zhenming等(1994)还报道,只要有3%~4%(体积)的部分熔融就能使得橄榄岩的流变强度发生巨大的降低,并将之归因于变形过程中全部的颗粒边界都遭到熔体的润湿。其实,Jin Zhenming等(1994)实验使用的Griggs装置对应力测量相当的不敏感(± 50 MPa, Karato and Weidner, 2008),加之高压胞中试样的具体形状未知以及温度极其不均匀,所以,无法准确地估计Griggs装置内变形试样的流变强度。

部分熔融体系的流变学性质和固—液相之间的成分差异亦具有较大的关系,例如,大洋中脊玄武岩浆与金属硫化物的熔体对橄榄岩变形及流变强度的影响就截然不同(图8)。在橄榄石+MORB的体系中,玄武岩浆既可作为高速扩散的通道又可造成应力的局部集中(Cooper and Kohlstedt, 1986; Cooper et al., 1989)。而在橄榄石+Fe—S熔体的

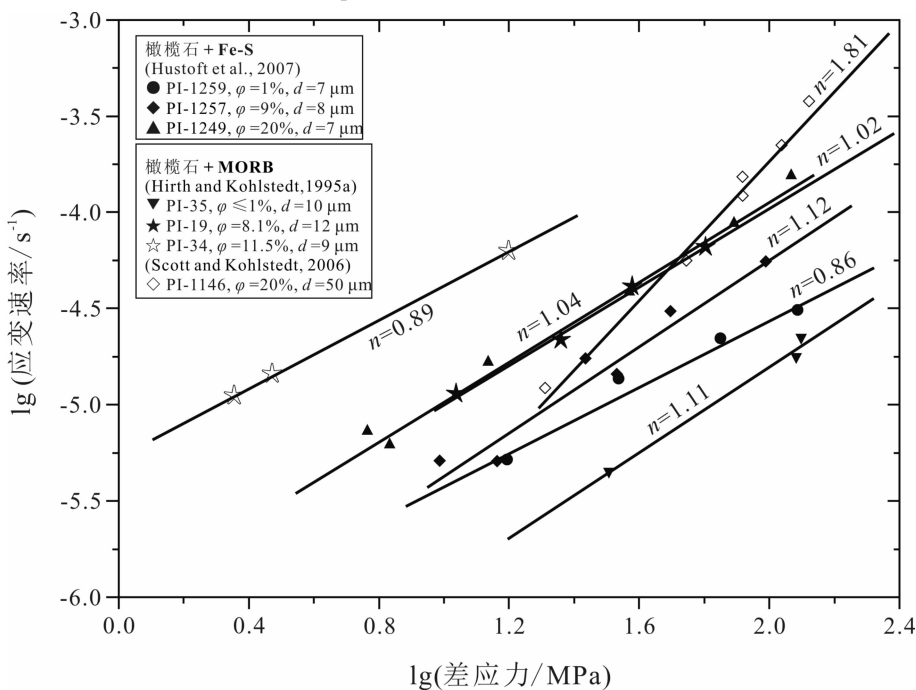


图8 具有不同熔体分数的橄榄石+Fe—S和橄榄石+大洋中脊玄武岩(MORB)样品在围压300 MPa和温度1523 K下的流变强度与应变速率的双对数投图(建图数据取自Hirth and Kohlstedt, 1995a; Scott and Kohlstedt, 2006; Hustoff et al., 2007). d 为橄榄石的颗粒径

Fig. 8 Log—log plots of differential stress versus strain rate for samples of olivine + Fe—S and olivine + MORB aggregates with different melt fractions deformed at a confining pressure of 300 MPa and at a temperature of 1523 K (Data from Hirth and Kohlstedt, 1995a; Scott and Kohlstedt, 2006; Hustoff et al., 2007). d : grain size of olivine

体系中,由于橄榄石在金属硫化物熔体内的溶解度很低,Fe—S的熔体就不可能为橄榄石组分的扩散提供高速有效的通道,因而 Fe—S 熔体对整体部分熔融岩石的流变强度的影响主要是由熔体导致橄榄石颗粒接触边界局部应力的增加所造成 (Hustoft et al., 2007)。显微构造分析表明,在上述两种体系中, MORB 和 Fe—S 熔体具有明显不同的润湿几何,其二面角分别为 38° 和 90° , Fe—S 熔体仅以孤立的熔体包的形式局限于三连点处,而 MORB 熔体即使在体积分数不大时也能润湿所有三连点及其粒棱。所以,在熔体体积分数相同的情况下,硅酸盐熔体的弱化效应要比金属熔体的大得多,这一认识对于理解超基性岩中金属成矿作用与围岩的韧性变形之间的关系具有重要的理论和实际意义。

在构造应力恒定的情况下,熔体体积分数的增加必将会提高部分熔融岩石的应变速率(图9)。在高应力和大应变时,熔体的拓扑结构从由界面张力控制转化成由应力控制,从而导致流变弱化 (Bussod and Christie, 1991; Zimmerman et al., 1999)。据 Mecklenburgh and Rutter (2003) 的实验

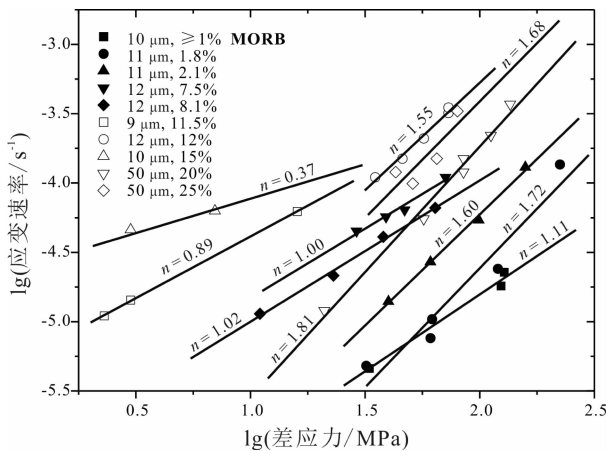


图9 具有不同熔体分数的橄榄石 + 大洋中脊玄武岩 (MORB) 样品在围压 300 MPa 和温度 1473 K (熔体分数为 2.1%、7.5% 和 8.1%) 与 1523 K 下的流变强度与应变速率的双对数投图 (建图数据取自 Hirth and Kohlstedt, 1995a; Mei Shenghua et al., 2002; Scott and Kohlstedt, 2006)

Fig. 9 Log—log plots of differential stress versus strain rate for samples of olivine + MORB aggregates with different melt fractions deformed at a confining pressure of 300 MPa and at a temperature of 1473 K (for melt fractions of 2.1%, 7.5%, and 8.1%) and 1523 K (Data from Hirth and Kohlstedt, 1995a; Mei Shenghua et al., 2002; Scott and Kohlstedt, 2006)

研究,花岗岩中的熔体体积分数从约 16% 增至 28%,导致其应变速率提高约 160 倍。在简单剪切变形过程中,熔体包的长轴还会逐渐沿着伸展剪切带作优选定向排列,导致进一步的几何弱化 (Zimmerman et al., 1999)。

如果熔体分数进一步增加并超过某一临界值 (即临界熔体分数 CMF: Critical Melt Fraction; Arzi, 1978), 熔体就会形成彼此贯通的网络,而之前连续的固态岩石则被分割成彼此孤立的岩块悬浮于弥漫的熔体之中,这样的悬浮系统相对于之前的固态岩石的强度大大降低 (图 10、图 11), 这样的弱化机制称为液化减弱 (Davidson et al., 1994; 嵇少丞等, 2008)。液化减弱的实质是岩石结构的转变,从强相支撑结构转变到弱相支撑结构 (Arzi, 1978; Ji Shaocheng and Xia Bin, 2002; Ji Shaocheng, 2004)。Arzi (1978) 发现, Westerly 花岗岩的粘度在约 12% 的熔体体积分数时就开始降低; 而 Goetze (1977) 认为只有熔体体积分数高达 25% ~ 30% 之后, 橄榄岩的流变强度才会大幅度的减低。后一种观点被 Scott and Kohlstedt (2006) 的实验证实 (图 10)。CMF 的概念及其地质意义已被许多作者讨论过 (Wickham, 1987; Bagdassarov and Dorfman, 1998; Barboza and Bergantz, 1998; Renner et al., 2000; Takeda and Obata, 2003)。Abe (1997)、Solomatov and Stevenson (1993) 和 Solomatov (2000) 据此探讨岩浆海冷却过程中流变强度的演化及其行星早期地幔演化的时间尺度等问题。

Ji Shaocheng (2004) 和嵇少丞等 (2006) 系统地分析了广义混合律在岩浆这样的悬浮体系的流变学中的应用。在晶体体积分数很小时, 矿物晶体彼此互不接触, 孤立地悬浮于连续的低粘度的熔体之中。在含高浓度、不规则形状晶体颗粒的悬浮液中, 晶粒相互碰撞、粘结成团, 部分流体填充到聚合团内的间隙中去, 从而减小了系统中流体的有效体积分数, 亦即增加了固体粒子的有效体积分数, 其结果增加了悬浮液的有效粘度。岩浆的流动强度 (σ_c) 由下式给出:

$$\sigma_c = \sigma_L \left(1 - \frac{V_s}{V_m} \right)^{(1/J)} \quad (1)$$

这里, σ_L 是纯熔体的流动强度, 下标 L 代表流体, J 是结构参数, V_s 是晶体的体积分数, V_m 是最大堆积密度。 V_m 取决于晶体的形状、粒径分布、形状优选定向 (SPO) 以及剪切应变量和剪切速率。在通常情况下, V_m 随颗粒的长/宽比 (aspect-ratio) 增加而减

少,有形状优选定向比没有形状优选定向具有更小的 V_m 值 (Saar et al., 2001)。若固体颗粒为等大球体,则 V_m 和球径无关,

$$V_m = \frac{\pi}{\sqrt{18}},$$

即 0.74, 此值对应于开普勒 (Johannes Kepler) 于 1611 年猜想中的等大球体的最大堆积密度 (Sziro, 2003; Ji Shaocheng, 2004)。通常 $V_m = 0.74$ 被看作是等大球体堆积密度在理论上不可逾越的最大值。Barnes 等 (1989) 为任意堆积体系推导出下列公式:

$$h = d \sqrt{\frac{1}{3\pi V_s} + \frac{5}{6}} \quad (2)$$

h 是相邻球心间距, d 是球径, V_s 是体系中球的体积分数。由于刚球不能相互嵌入或重叠, 所以 h 的最小取值应为 d , 此时 V_s 有最大值 $2/\pi$, 即 0.637, 此值就是任意最紧密堆积密度。将等大刚球置入已知体积刚性容器的实验也证明任意紧密堆积的密度介于 0.601 和 0.637 之间 (Allen, 1984; Rogers et al., 1994)。前人的理论分析和计算机模拟得出等大球体的任意紧密堆积密度约为 0.64 (Jaeger and Nagel, 1992; Torquato et al., 2000; Jalali and Li Mo, 2004)。由此可见, 任意紧密堆积密度远小于立方面心或六方最紧密堆积密度。 V_m 的下限值应

该是 0.555 (Onoda and Liniger, 1990), 此值对应于等大球体的任意疏松堆积。所以, 等大球体悬浮体系液态行为向固态行为的转变应出现在 $V_s = 0.555$ 和 $V_s = 0.640$ 之间。对于自然界含非球状矿物晶体的岩浆, V_m 一般约为 0.60 (Marsh, 1981)。此外, V_m 随剪切应变量的增加而增加, 这是因为矿物晶体的形状优选定向程度随剪应变的增加而增强 (Tsenoglou, 1990; Jones et al., 1991)。 V_m 还取决于颗粒径分布的幅度。小颗粒会填进大颗粒间的空隙, 从而腾出部分空间来。粒径呈单峰分布的颗粒要比多峰分布的颗粒更易形成应力支撑格架固体结构。在相同的条件下, 粒径呈多峰分布的体系的相对粘度比粒径呈单峰分布的体系能小一两个数量级 (图 12, Poslinski et al., 1988), 这就是固-液体体系触变性 (Thixotropy) 的科学原理 (Barnes, 1997)。 V_m 值还随颗粒表面糙度 (Roughness) 增加而减小 (Kitano et al., 1981)。

理论上说, 临界熔体分数 $CMF = 1 - V_m$ 。所以。当 V_s 达到或超过 V_m 时, 固体颗粒相互连接形成应力支撑格架, 存在于颗粒间隙中的流体分量不足, 不再能对粒间运动起润滑作用。此时, 体系从流体行为转变为固态行为。所以, 式 (1) 仅适用于 $V_s < V_m$ 或 $V_L > V_{CMF}$ 的岩浆体系。

所有现存的各种实际悬浮液的实验数据都说明广义混合律中的 $J = -0.5$ (Ji Shaocheng, 2004; 嵇少丞等, 2006)。图 11 表示花岗岩-熔体体系的流变强度的演化曲线, 其中 AF 线就是根据式 (1) 采用 $J = -0.5$, $V_m = 0.64$ (即等大球体的任意紧密堆积密度) 算出来的。图中的临界熔体分数 $CMF = 0.36$, 会随着晶体的长/宽比增加或粒度分布的单峰性增加而增加, 即向右边移动。

对于 $V_s > V_m$ 或 $V_L < CMF$ 的部分熔融体系, 由于熔体的强度相对于固体来说完全可以忽略不计, 这样的体系可以近似地看成是孔隙材料 (Tharp, 1983; Hirth and Kohlstedt, 1995a), 其中固体

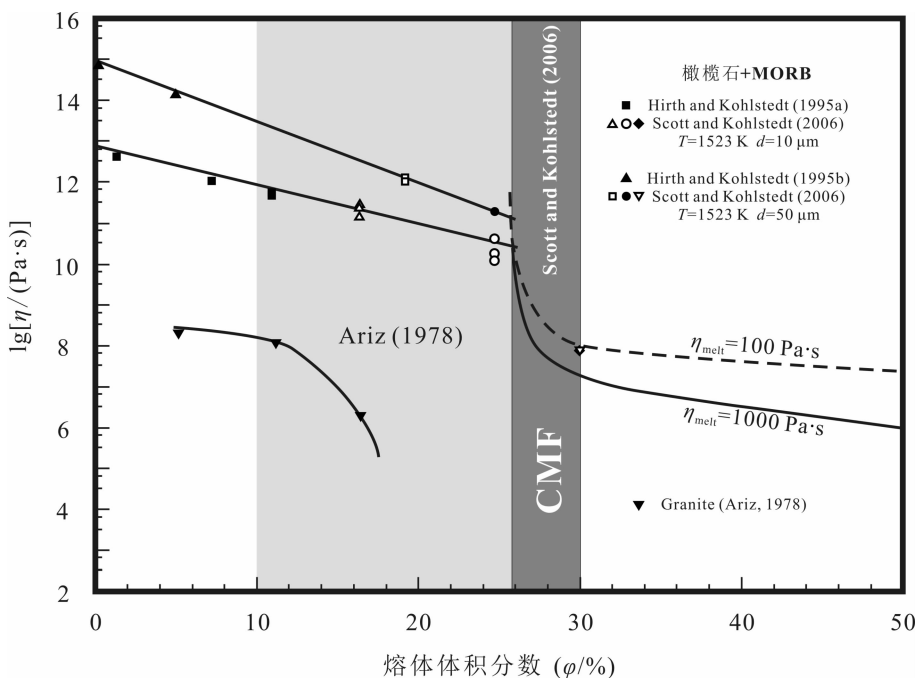


图 10 部分熔融岩石的粘度与熔体分数的半对数投图

Fig. 10 Semi-log plots of viscosity vs. melt fraction for partially molten rocks

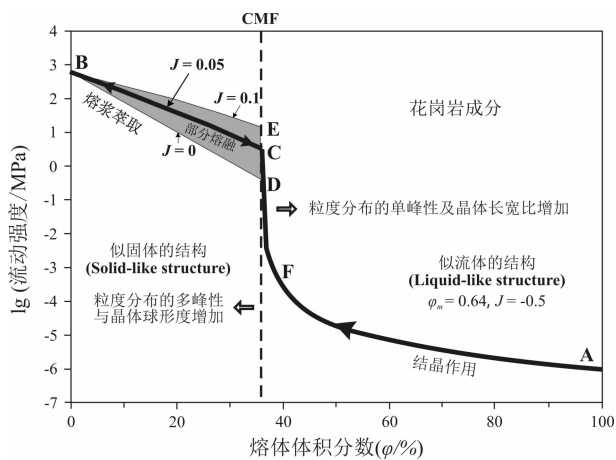


图 11 由广义混合律描述的部分熔融花岗岩的流动强度随熔体体积分数的变化规律

Fig. 11 $\lg(\text{flow strength})$ as a function of melt volume fraction (φ) for granitic composition, predicted by the generalized mixture rule

构成连续的应力支撑格架。孔隙材料的强度可以由下式估算 (Ji Shaocheng, 2004; 嵇少丞等, 2006):

$$\sigma_c = \sigma_s \left(1 - \frac{p}{p_c}\right)^{(1/J)} \quad (3)$$

其中, σ_s 和 σ_c 分别表示纯固体无孔隙材料的强度和孔隙材料的强度, p 是孔隙材料的孔隙度, p_c 是材料中最大可能的孔隙度, 只要材料的孔隙度超过这个临界值 p_c , 连续固体支撑格架立马遭到破坏, 岩石的整体有效强度随之消失。对于含等大球状并作

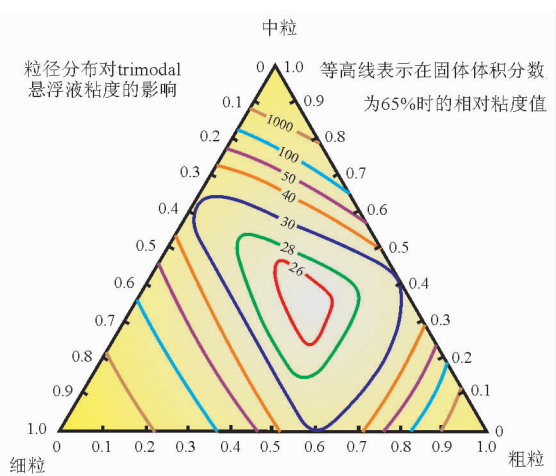


图 12 含 65% (体积分数) 固体球的悬浮液的粘度随粗、中、细三种不同大小固体球比例的变化

Fig. 12 Effects of particle size distribution on trimodal suspension viscosity. Contours show values of the relative viscosity at the solid volume fraction of 65%

正方、立方或菱面体几何排列的孔隙材料, Knudsen (1959) 计算得到它们的 p_c 值分别是 0.476、0.397 和 0.26。对粉末材料如戈壁滩上的沙子或干燥土壤, p_c 值对应于压实之前的散堆孔隙度 (Apparent porosity)。 p_c 取决于孔隙的形状、排列、分布、组合及其贯通性。对应于球状的和任意非球状孔隙的孤立、不贯通的孔隙材料, J 值分别为 0.5 和 0.25 (Ji Shaocheng et al., 2006)。

嵇少丞等 (2006) 分析了众多的实验数据后得出, 部分熔融基性和超基性岩的 J 值约为 0.005 (图 13a), 而部分熔融花岗岩的 J 值主要在 0 到 0.1 之间变化 (图 13b), 其平均值为 0.05 (对应于图 11 中的 BC 线)。显而易见, 上述的孔隙模型 ($J = 0.5$ 对应于球状孔隙, $J = 0.25$ 对应于任意非球状孔隙) 明显地低估了熔体对部分熔融岩石流动强度的影响。虽然相对晶体来说, 无论孔隙还是熔体的强度都可忽略不计, 然而两者对材料整体流动强度的影响却大为不同, 其差别在于晶体和孔隙之间没有化学反应 (图 14), 而熔体的存在却能改变材料的变形机制, 例如熔体可以润滑粒间边界, 促进颗粒边界滑移, 还可以增强溶解—沉淀作用, 为 Coble 蠕变提供可持续性的晶界扩散的捷径 (如, Hirth and Kohlstedt, 1995a), 这就是熔体比孔隙对材料的强度降低影响更为显著的原因。

随着体系内熔体润湿程度的逐渐增加, 多晶集合体内部的变形机制也会发生相应的改变 (Pharr and Ashby, 1983; Beeman and Kohlstedt, 1993), 往往会从位错蠕变转变为粒间熔体促进的扩散蠕变, 前者与粒径无关, 而后者对粒径十分敏感。现已发现, 上述变形机制的转变出现在部分熔融的花岗岩 (Dell' Angelo et al., 1987; Dell' Angelo and Tullis, 1988; Tullis et al., 1996)、钠长岩 (嵇少丞, 1988)、钙长岩 (Dimanov et al., 2000) 和橄榄岩 (Hirth and Kohlstedt, 1995a, b) 中。

此外, 在含变价元素的固—熔体的体系中, 熔致弱化作用 (Melt-induced weakening) 还受岩石变形的化学环境特别是氧逸度的影响, 这是因为化学环境亦会影响熔体的几何形态与分布, 进而影响岩石的流变强度 (周永胜等, 2008)。Kohlstedt and Zimmerman (1996) 在对橄榄石 + 3% (体积分数) MORB 的变形实验中发现, 应变速率随着氧逸度增加而增加 (图 15), 这与前人 (Bai Quan et al., 1991; Beeman and Kohlstedt, 1993) 对单相橄榄石的研究结果是一致的。具体地说, 除了在 1400°C 时

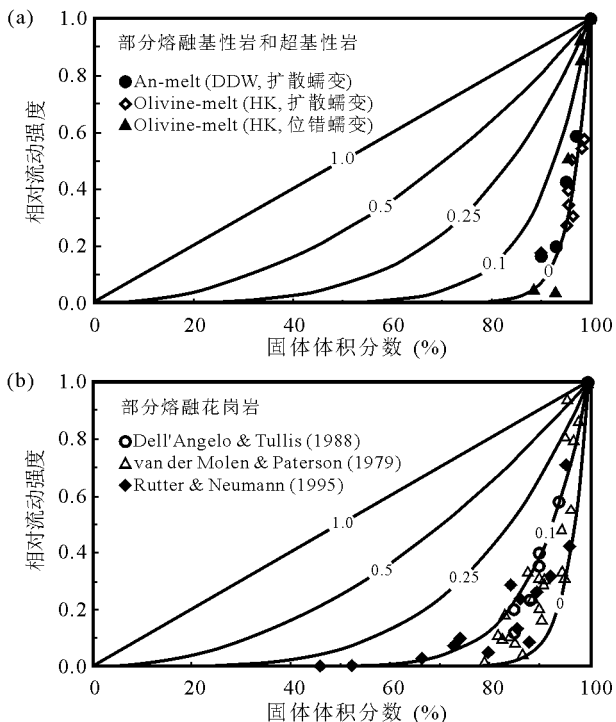


图 13 部分熔融岩石的流动强度随固体体积分数的变化。理论曲线上的标数为广义混合律的构造参数 J 值：(a) 部分熔融超基性岩，数据取自 Dimanov 等 (1998) (DDW) 和 Hirth and Kohlstedt (1995a, b) (HK)；(b) 部分熔融花岗岩

Fig. 13 Normalized flow strength vs. solid volume fraction for partially molten rocks. Theoretical curves labeled according to J values of the generalized mixture rule: (a) Partially molten ultramafic rocks. The data from Dimanov et al. (1998) (DDW) and Hirth and Kohlstedt (1995a, b) (HK); (b) Partially molten granites

应变速率—氧逸度的变化线的斜率稍缓以外，其他温度时变化线的斜率都非常相似 (图 15)。Minarik 等 (1996) 和 Gaetani and Grove (1999) 实验研究了硅酸盐 + Fe—S 的熔体体系，发现熔体的二面角随氧逸度增加而减小，在氧逸度高的环境中，橄榄石 + Fe—S 体系中熔体二面角小于 60° ，而在低氧逸度环境中则为 90° (Hustoft et al., 2007)。据此推测，在氧逸度高的环境如火星地幔中 (Rubie et al., 2004)，金属硫化物熔体导致的岩石流变强度的降低会更加显著。此外，压力升高亦会导致橄榄岩中 Fe—S 熔体的二面角减小，达到甚至低于 60° ，从而导致整个体系流变强度的剧烈下降 (Minarik et al., 1996; Shannon and Agee, 1998; Takafuji et al., 2004)，这种现象与橄榄石—水 (Mibe et al., 1998;

1999; Yoshino et al., 2007) 和橄榄石—MORB 体系 (Yoshino et al., 2009) 的相似，说明压力的确是影响部分熔融岩石流变学性质的重要因素之一。

有趣的是，在熔融分数很小的时候，部分熔融还有可能提高橄榄岩的强度，这是因为水在玄武岩熔体中的溶解度 (Sisson and Grove, 1993) 要远远大

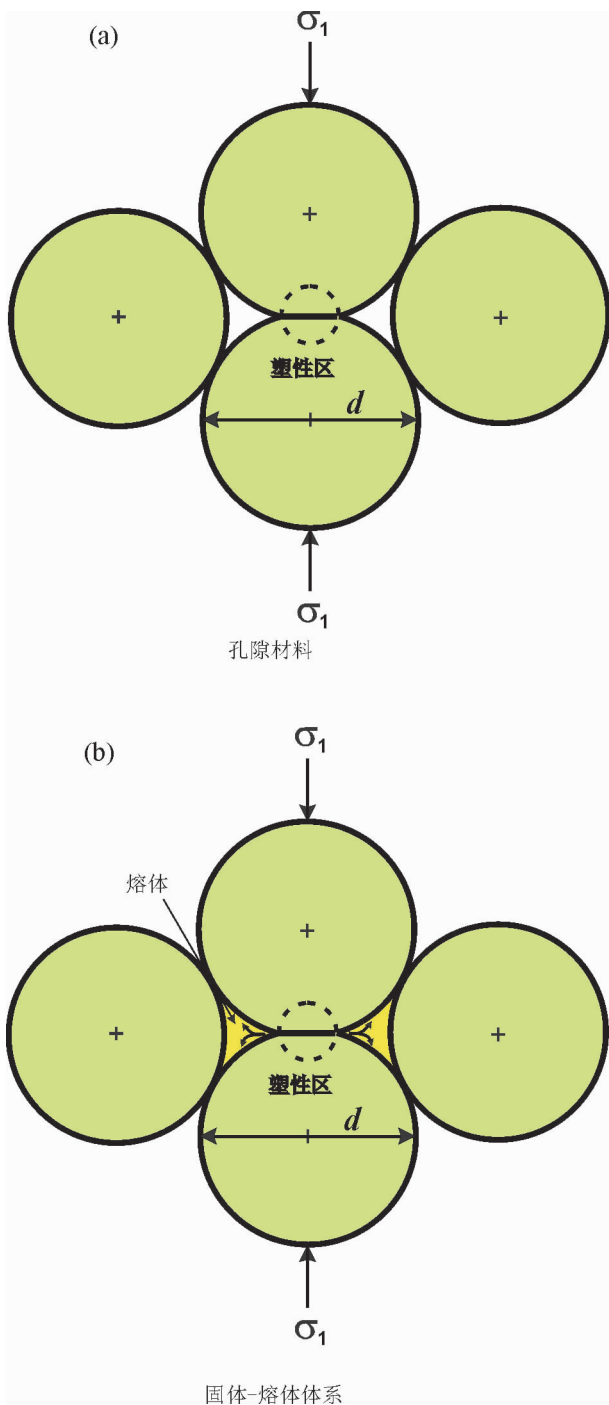


图 14 无充填孔隙材料与固体—流体体系变形的差别
Fig. 14 Differences between deformations of a porous material and a solid—liquid system

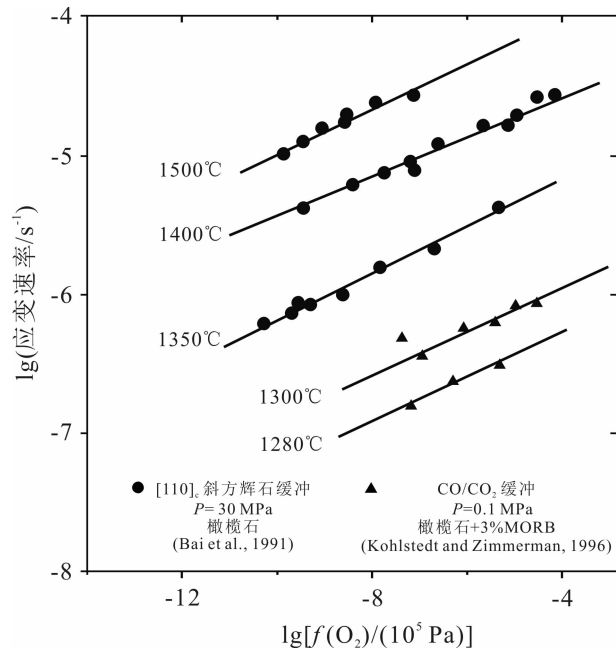


图 15 橄榄石单晶与橄榄石 + 3% (体积分数) MORB 在低压和各种温度下的应变速率与氧逸度的关系曲线
 Fig. 15 Strain rate versus oxygen fugacity for samples of olivine single crystals and polycrystalline olivine plus 3 vol% MORB deformed at low pressure and various temperatures

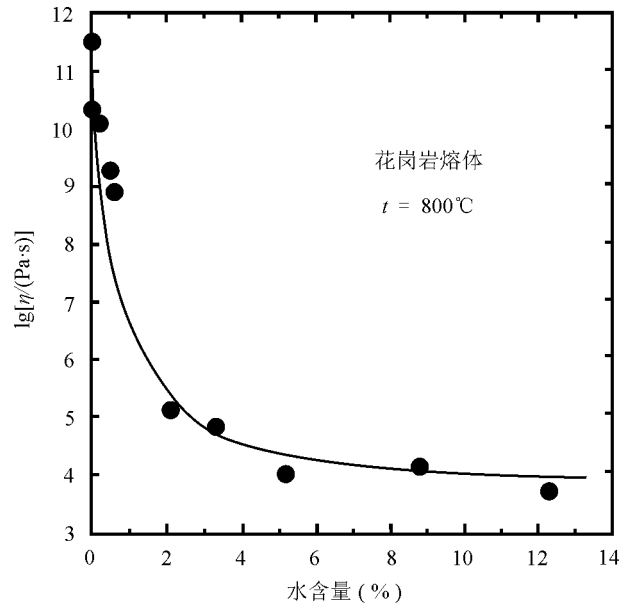


图 16 800°C 时花岗岩熔体的粘度与水含量的关系。建图数据来自于 Friedman 等 (1963), Shaw (1963), Burnham (1964), Persikov (1991), Baker (1996) and Clemens and Petford (1999)
 Fig. 16 Viscosity of granitic melts as a function of the weight percent of dissolved water. Data measured by Friedman et al. (1963), Shaw (1963), Burnham (1964), Persikov (1991), and Baker (1996)

于其在橄榄石或辉石中的溶解度 (Bai Quan, 1992; Kohlstedt et al., 1996)。例如, Grant 等 (2007) 发现水在橄榄石、斜方辉石和熔体之间的分配比例是 10: 33: 100000。由此可见, 在部分熔融过程中, 橄榄石和辉石晶格中的水优先扩散进入熔体 (Karato, 1986; Hirth and Kohlstedt, 1996; Grant et al., 2007; 嵇少丞等, 2008), 橄榄石和辉石晶体变干了。大家知道, 水含量的增加不仅会有效地降低岩石的固相线 (熔点) 和熔体的粘度 (图 16), 而且会减弱硅酸盐矿物的 Si—O 键, 引起水致弱化作用 (Griggs and Blacic, 1965; Griggs, 1967; Chopra and Paterson, 1984; Paterson, 1990)。所以, 只要部分熔融造成橄榄石和辉石晶体变干了, 也就会理所当然地导致岩石流变强度的提高。这一认识最早由 Karato (1986) 提出, 随后得到了 Hirth and Kohlstedt (1995a, b) 等实验研究的支持, 而且部分熔融强化岩石的认识在斜长岩的流变实验中亦得到了证实 (Huang Jianguo et al., 2001)。Hirth and Kohlstedt (1996) 将这一理念拓展并据此建立了洋中脊之下地幔岩石的流变学模型。然而, 这一模型没有考虑熔致弱化与脱水强化之间的竞争效应 (Hirth and Kohlstedt, 1996; Karato, 2008, 2010), 故其有效性

尚值得怀疑, 能否直接用到洋中脊动力学的模拟中去, 更无从定论 (Ito et al., 1996, 1999; Ito, 2001)。Karato (2008) 详细探讨了熔致弱化与脱水强化之间的竞争效应, 并给出了部分熔融对岩石流变强度的净效应表达式:

$$\dot{\epsilon} = \dot{\epsilon}_0 \cdot F_1(\varphi_1) \cdot F_2(\varphi_2)$$

这里的 $\dot{\epsilon}$ 是部分熔融材料的应变速率, $\dot{\epsilon}_0$ 是部分熔融前的应变速率; $F_1(\varphi_1)$ 代表脱水强化效应, 为熔融程度 φ_1 的函数; $F_2(\varphi_2)$ 代表熔致弱化效应, 取决于熔体分数 φ_2 和熔体包几何 (如, θ)。一般情况下, $F_1(\varphi_1) < 1$ 、 $F_2(\varphi_2) > 1$, 因而部分熔融的净效应取决于上述两种效应的相互竞争, 即 $F_1 \cdot F_2 > 1$ 或者 $F_1 \cdot F_2 < 1$ 。详细的讨论可参阅 Karato (2008) 第 19 章。

在部分熔融的理论研究方面, Raj and Ashby (1971)、Cooper and Kohlstedt (1986)、Cooper 等 (1989)、Takei (1998)、Hier-Majumder (2004)、Karato (2008) 和 Takei and Holtzman (2009a, b, c) 等已经做出了一些工作, 其中 Cooper and Kohlstedt (1986) 和 Cooper 等 (1989) 提出在橄榄岩中玄武

岩熔体在三连点和粒棱作各向同性分布,基于CK模型,他们预测了熔致弱化效应。然而,随后的实验研究(Hirth and Kohlstedt, 1995a, b)表明,CK模型低估了熔体分数大于5%时熔体对应变速率的影响,这是因为在较高熔体分数时橄榄石—熔体界面能的各向异性会导致熔体润湿大部分颗粒边界,从而提高岩石的变形速率(Kohlstedt, 2002; Zimmerman and Kohlstedt, 2004)。最近,基于CK模型,Takei and Holtzman (2009a, b, c)兼顾了颗粒之间的质量守恒、力学平衡以及反应与扩散动力学平衡,提出了关于部分熔融岩石颗粒边界扩散蠕变的一种新模型,据此他们对颗粒边界驱动扩散流动的应力的分布进行了更为详细的分析和计算,但是他们的模型所预言的低熔体分数时的熔致弱化要比实验测量(如, Hirth and Kohlstedt, 1995a, b; Kohlstedt, 2002)得出来的稍大。陡的应力梯度分布可能是Takei and Holtzman (2009a, b, c)的模型预言出高应变速率的主要原因,因为陡的应力梯度能够增强和促进物质的扩散(Raj, 1975)。然而,这种强化变形仅是瞬时的,到了稳态时,应力分布也就不再具有陡的梯度,此时的蠕变速率就与传统模型(如Raj and Ashby, 1971)的结果非常一致了。造成Takei and Holtzman (2009a, b, c)的模拟结果与实验结果(如, Kohlstedt, 2002)之间差异的原因还有待进一步确定。

4 结论

在静态条件下,熔体二面角除了与温压条件、晶体形态、固—固相以及固—液相的界面自由能等因素相关外,还强烈取决于熔体的化学组分。例如,在盐—水体系中,熔体二面角就非常小,几乎为零(Urai, 1983; Urai et al., 1986),因此,不能盲目地将盐—水体系的研究结论直接应用于由铝硅酸盐或硅酸盐矿物组成的上地幔岩石或地壳岩石。橄榄岩中的熔体二面角随着静水压力的升高也会有所减小(Yoshino et al., 2007, 2009)。熔体一般会优先润湿造岩矿物的低指数晶面(Waff and Faul, 1992; Faul et al., 1994)。在有差应力存在的动态条件下,熔体二面角往往还受变形的化学环境(如,氧逸度)的影响。一般原则是,在静态条件下,熔体迁移的主要驱动力是熔体表面张力;而在动态条件下,熔体的几何形态与分布既会受到化学扩散又会受到差应力的控制,因而各具各向异性的特征。

从上世纪80年代起,许多学者对含熔体的主要

造岩矿物多晶集合体进行了大量的流变学研究,其中对部分熔融橄榄岩的低压(< 500 MPa)实验研究(Hirth and Kohlstedt, 1995a, b; Kohlstedt and Zimmerman, 1996; Zimmerman, 1999; Mei Shenghua et al., 2002; Scott and Kohlstedt, 2006)与理论研究(Cooper and Kohlstedt, 1986; Cooper et al., 1989; Takei and Holtzman, 2009a, b, c)都表明,在低熔体体积分数($< 约5%$)时,部分熔融对橄榄岩只具中等的弱化效应。然而,据Jin Zhenming等(1994)报道,在低的熔体分数和中等围压(约1.5 GPa)条件下,差应力导致熔体几乎完全润湿橄榄石的颗粒边界,导致强烈的弱化。Mibe等(1999)和最近Yoshino等(2007)的静态高压(> 2.0 GPa)实验证明,流体的二面角会随着压力的升高而减小,在约7.5 GPa时他们发现橄榄石颗粒边界被流体完全润湿(Yoshino et al., 2007)。但是,他们的实验是在静水压力的条件下进行的,因而无法查明该体系流变学强度随熔体二面角的变化规律。在有差应力和大变形的条件下,熔体的拓扑结构从受界面张力控制转化成由应力控制,流变弱化更加显著。随着熔体分数的增加,熔体会分离出来,汇聚到伸展剪切带,造成熔体在岩石中形成优选定向(MPO, Holtzman et al., 2003a; Scott and Kohlstedt, 2006; Holtzman and Kohlstedt, 2007),与此同时,固体岩石的主导变形机制往往由位错蠕变转变成扩散蠕变(嵇少丞, 1988; Hirth and Kohlstedt, 1995a, b; Daines and Kohlstedt, 1997)。

致谢:感谢两位审稿人提出的建设性修改意见。

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Lett. , 26(10) : 1505 ~ 1508.

Rheology of Partially Molten Rocks: a State-of-the-art Overview

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Abstract: This paper provides a state-of-the-art overview on the melt morphology and distribution in partially molten rocks under hydrostatic and dynamic conditions with special focuses on the influence of partial melting on rheological behavior of peridotite and granite. Partial melting controls most of the important geological processes (e. g. , formation of the crust, compositional evolution of the mantle) and the physical properties of deep crust and upper mantle (e. g. , electrical conductivity, anelasticity, seismic wave velocities, and permeability). The understanding of the rheological properties of partially molten rocks is thus essential for modeling precisely large-scale geological processes including mechanical coupling between the crust and lithospheric mantle, convection of asthenosphere, plate tectonics, channel flow of middle to lower crust, mountain-building mechanisms. Significant progress in both experimental and theoretical studies occurred during the last three decades and has promoted our understanding of the rheological properties of partially molten rocks. The following consensus has been reached: under hydrostatic conditions, melt whose volume fraction is less than $\sim 2\%$ is generally restricted to the triple junctions or along grain edges. The degree of melt wetting increases with increasing pressure, temperature and melt fraction. Under non-hydrostatic conditions (either coaxial compression or simple shear), however, melt occurs predominantly along extensional shear bands aligned at angles of about $15^\circ \sim 30^\circ$ to the maximum principal compressive stress. The presence of $< 5\%$ melt results hence in only a modest rheological weakening due to the heterogeneous distribution of melt. With increasing the melt fraction, the partially molten rocks are significantly weakened, which leads to deformation-induced melt segregation or extraction.

Key words: Partial melting; Melt morphology and distribution; Dihedral angle; Rheology; Large shear strain; Geodynamics