

Institut de Physique du Globe de Paris
Ecole doctorale des Sciences de la Terre

Thèse de Doctorat

pour l'obtention du titre de

Docteur en Science

de l'Institut de Physique du Globe de Paris

Spécialité : GEOPHYSIQUE

Soutenue par

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Magmatisme intrusif sur les planètes telluriques

Équipe PLANÉTOLOGIE ET SCIENCES SPATIALES,

Défendue le 5 Décembre, 2013.

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Remerciements

Je remercie marion rouault mae sans qui tout ceci n'aurait jamais vu le jour
;)

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Rsum de la probltique et rsltats principaux

Part I

Dynamique des magmas magma à faible profondeur

CHAPTER 1

Magmatisme intrusif

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1.1 Formation, transport et stockage des magmas

1.1.1 Formation

La majorité des magmas sont formés par fusion partielle des roches du manteau supérieur. Dans les conditions normales de pression, la température du manteau supérieur n'est pourtant pas suffisante pour provoquer leur fusion (Figure 1.1) et d'autres mécanismes sont nécessaires pour amener les roches du manteau à croiser leur liquidus. Au niveau des dorsales en contexte océanique ou des rifts en contexte continental ou encore au sein des panaches mantelliques, la fusion partielle est ainsi causée par décompression (Figure

1.1 b). Au niveau des zones de subduction, les mécanismes mis en jeux sont plus complexes et font intervenir la déshydratation par chauffage des roches, la migration des fluides abaissant le liquidus et ainsi provoquant la fusion des roches alentour (Figure 1.1 c).

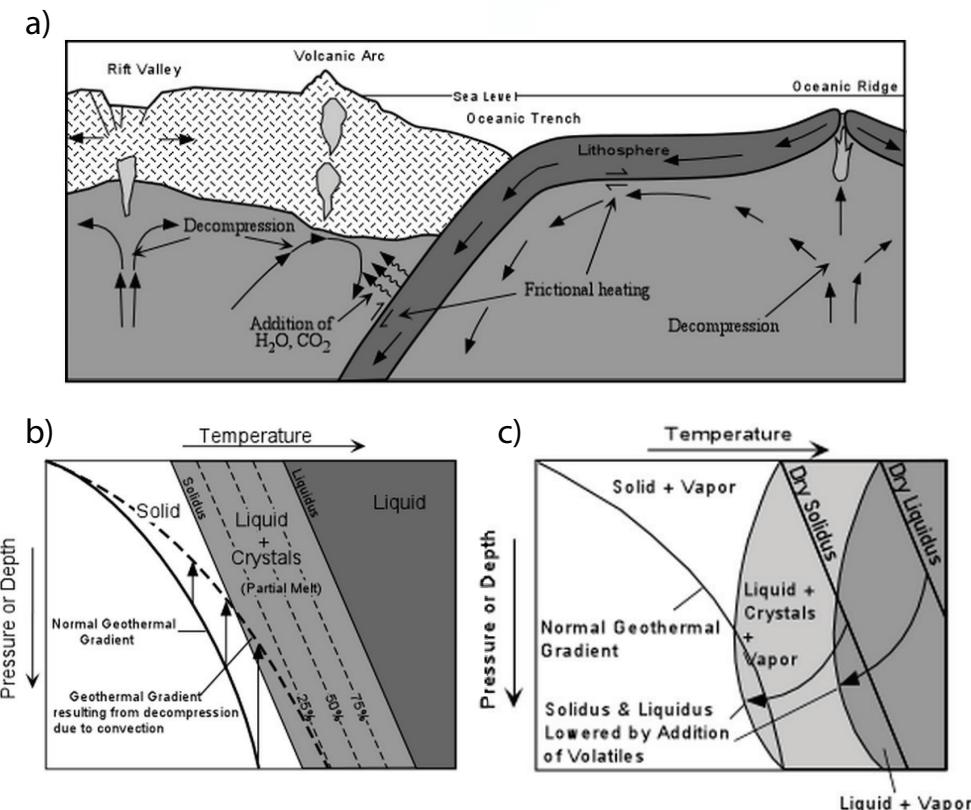


Figure 1.1: a) Lien entre le magmatisme et la tectonique des plaques: production de magma par fusion partielle par décompression au niveau des dorsales océaniques ou des rifts en contexte continental ou par addition de volatiles au niveau des zones de subduction. Schéma du diagramme de phase des roches du manteau supérieur dans deux contextes différents: a) dorsale océanique ou panache mantellique, b) zone de subduction.

1.1.2 Transport

Les liquides de fusions ainsi formés sont moins denses que les roches solides alentour et s'élèvent donc, par compaction et percolation au travers de la matrice mantellique ([McKenzie, 1984, 1985](#)). Le magma, liquide de fusion +

cristaux, s'accumule ensuite au sein de chenaux, i.e. de dykes ou le long de faille préexistantes pour remonter rapidement vers les couches superficielles de la croûte (*Lister and Kerr, 1991; Clemens and Mawer, 1992; Petford et al., 1993; Rubin, 1995*). En effet, bien que l'idée du magma remontant lentement au sein de gros volume diapirique soit encore parfois invoquée au sein de la base ductile de la croûte (*Weinberg and Podladchikov, 1994; Weinberg, 1996*), le transport rapide du magma au sein des dykes permet de résoudre de nombreux problèmes, thermiques et mécaniques, associés à la remonté diapirique de gros volume de magma au sein des parties supérieures fragiles de la croûte invoquée historiquement (*Miller and Paterson, 1999*).

1.1.3 Stockage

Les travaux de *Walker (1989)* ont montré que les magmas remontent jusqu'à rencontrer leur zone de flottabilité neutre, une région où la densité de la roche encaissante est proche de celle du magma lui-même. En effet, au-dessus de cette couche, le magma est plus dense que la roche encaissante et sa flottabilité l'entraîne vers le bas. De nombreux travaux, tant théoriques (*Lister and Kerr, 1991; Petford et al., 1993; Rubin, 1995*) que expérimentaux (*Taisne and Tait, 2009; Taisne et al., 2011*) ont en effet depuis montré que l'ascension d'un dyke était contrôlée par la différence de densité entre la tête de celui-ci et la roche encaissante. Lorsque le dyke entre dans une région de densité inférieure, la surpression induite peut, sous certaines conditions, conduire à l'étalement du magma au niveau de la base de la région de plus forte densité permettant ainsi ainsi la formation de réservoir magmatique sous forme d'intrusion magmatique au sein de la croûte (*Taisne et al., 2011*).

Plus récemment, d'autres études ont montré que les contrastes de rigidité entre les différentes couches crustales pourraient aussi jouer un rôle non négligeable sur l'arrêt de l'ascension des dykes (*Menand, 2011*). En effet, des expériences réalisées par *Kavanagh et al. (2006)* ont montré que la propagation d'un dyke peut être arrêtée quand celui-ci rencontre une interface qui sépare un milieu plus rigide surplombant un milieu moins rigide (Figure 1.2). Le dyke arrête ainsi son ascension verticale et s'étale horizontalement juste en dessous de la couche de rigidité plus élevée. Ce mécanisme serait d'autant plus efficace que le contraste de rigidité est important (*Kavanagh et al., 2006*).

Finalement, les contraintes, locales ou globales, peuvent aussi dévier la trajectoire d'un dyke et influencer les trajets des magmas au sein de la croûte. En effet, des études ont montré que les chenaux par lesquels se propage le magma tendent à s'orienter perpendiculairement aux contraintes de compressions (*Anderson, 1951*). Les dykes ont donc tendance à exister dans des situations où les contraintes de compressions sont horizontales et donc à s'étaler quand le

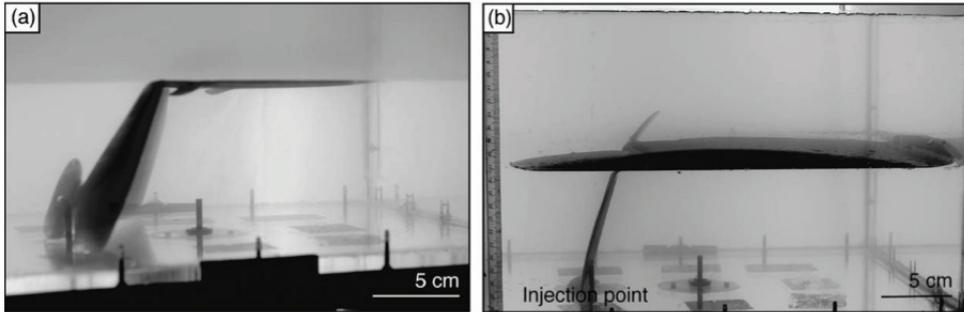


Figure 1.2: a) Photographie de deux des expériences réalisées par *Kavanagh et al.* (2006) sur le comportement d'un dyke à l'interface entre deux milieux de rigidités différentes. a) Le contraste de rigidité est très important et le dyke s'étale sous la couche de rigidité importante. b) Le contraste de rigidité est plus faible et, tout en s'étalant en dessous de la couche de rigidité supérieure, le dyke continue sa progression dans le milieu plus rigide.

champ de contrainte évolue d'une contrainte de compression horizontale à vertical comme c'est le cas par exemple au niveau des édifices volcaniques (*Pinel and Jaupart*, 2000, 2004; *Roman and Jaupart*, 2014).

Cependant, si ces différents facteurs jouent sûrement tous un rôle sur le contrôle des trajets des magmas au sein de la croûte, la densité relative du magma et de la roche encaissant et donc l'existence d'une zone de flottabilité neutre est certainement le facteur déterminant dans la mise en place d'intrusions magmatiques. Le magmatisme intrusif, et donc la question du stockage des magmas, est donc de manière générale étroitement lié à la structure en densité de la croûte elle-même.

1.2 Importance et multiples visages du magmatisme intrusif

1.2.1 Magmatisme intrusif sur Terre

Sur Terre, la composition de la croûte, et donc sa densité, est bimodale. Au niveau des océans, la croûte océanique présente une nature essentiellement basaltique avec une densité moyenne proche de 2900 kg m^{-3} . Elle est formée continuellement au niveau des dorsales océaniques et recyclée, environ 200 Ma d'année plus tard au niveau des zones de subduction. Elle est épaisse en moyenne de 6 km et couvre à elle seule 70% de la surface du globe. Au contraire, la croûte continentale, qui occupe les 30% restants, présente une

composition plus évoluée et globalement andésitique avec une densité moyenne plus proche de 2700 kg m^{-3} . Elle est beaucoup plus vielle que la croûte océanique et est âgée en moyenne de 2.5 Ga, avec certaines roches ayant été daté jusqu'à 4 Ga d'années. Elle est aussi beaucoup plus épaisse que la croûte continentale; son épaisseur moyenne est de 35 km et peut excéder les 70 kilomètres sous certaines chaînes de montagnes comme l'Himalaya.

De par sa densité relativement basse, en particulier au niveau des continents, la croûte constitue un filtre efficace à la remontée des magmas en surface qui sont donc préférentiellement stockés en profondeur sous forme d'intrusions magmatiques. *Crisp* (1984) et *White et al.* (2006) estiment en effet que les volumes de lave extrudée à la surface sont relativement faibles en comparaison aux volumes mis en place au sein de la croûte terrestre, i.e. 5 fois plus faible en contexte océanique et jusqu'à 10 fois plus faible en contexte continental. Le magmatisme intrusif apparaît donc comme un processus essentiel dans la formation de la croûte. Sur Terre, les mouvements tectoniques en son sein ainsi que l'erosion ont permis d'exposer certaines de ces intrusions à la surface. En plus de leur taille, qui peut varier de quelques mètres à des centaines de kilomètres, la morphologie de ces intrusions montre une grande variabilité.

Les batholites sont de loin les plus imposants représentants de cette famille d'intrusion magmatique se mettant en place au sein de la partie fragile de la croûte. Ils peuvent atteindre jusqu'à quelques kilomètres d'épaisseur et s'étendre sur des centaines de kilomètres. Par exemple, le batholite de Sierra Nevada est une intrusion granitique qui s'étend sur presque la totalité de la Sierra Nevada en Californie. Des données géochronologiques sur certain de ces batholites ont montré que leur mise en place peut s'échelonner sur quelques millions d'années, un temps beaucoup plus grand que les temps raisonnables pour le refroidissement d'une chambre magmatique dans la partie fragile de la croûte (*Glazner et al.*, 2004). En effet, il est maintenant clair que la mise en place de ces gigantesques volumes de magmas se fait par incrément successifs de petits volumes de magma se solidifiant lors de leur mise en place sur de longues échelles de temps 10^5 to 10^6 années (*Petford et al.*, 2000; *Glazner et al.*, 2004). Dans cette thèse, on va donc se focaliser sur les mécanismes de formations et de mise en place de volumes intermédiaires de magma dans la partie fragile de la croûte continentale, à des profondeurs inférieures à 10 km.

Des études géologiques de terrain ont montré la présence de quatre grandes familles d'intrusions magmatiques de taille intermédiaire à faible profondeur. Deux de ces familles, les dykes et les bysmalites, sont discordantes, c'est-à-dire qu'elles se mettent en place perpendiculairement à la stratification naturelle de l'encaissant et deux autres, les sills et les laccolites, sont concordantes, i.e. elles se mettent en place parallèlement aux couches géologiques.

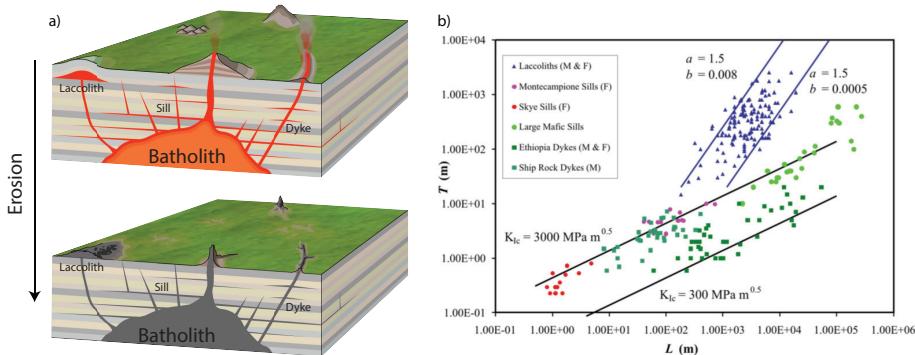


Figure 1.3: a) Différentes formes du magmatisme intrusif: batholite, dyke, sill et laccolite. Dimensions typiques pour des laccolites, dyke et sill de composition et d'origines différentes repris de *Cruden et al.* (2012).

- Les dykes, par lesquels remontent le magma à travers la lithosphère sont discordants et caractérisé par de faibles rapports d'aspects (Figure 1.3, 1.4 a). Leur épaisseur peut varier de quelques mètres à quelques centaines de mètres d'épaisseur (*Walker*, 1989; *Rubin*, 1995), cependant, l'épaisseur moyenne est de quelques dizaines de mètres. Les dykes de compositions felsiques sont généralement plus épais et moins longs que leurs équivalents mafiques (*Rubin*, 1995).
- Les sills, à la différence des dykes, sont concordants (Figure 1.3, 1.4 b,f). Ils se mettent en place le long de discontinuités ou de failles préexistantes, à la jonction entre deux couches sédimentaires par exemple. Les sills aux dimensions les plus importantes répertoriés sont mafiques et peuvent atteindre jusqu'à 100 km sur des épaisseurs de presque 1 km (*Cruden et al.*, 2012). Leurs homologues felsiques, plus rares, sont souvent de dimension plus faible.
- Les laccolites ont été décrit premièrement par *Gilbert* (1877) suite à son étude géologique des Henry Mountains, dans l'Utah aux États-Unis (Figure 1.4 c, d, e). Ils se mettent en place principalement par flexion des couches sédimentaires sus-jacentes, ce qui leur donne une forme de dôme caractéristique. Certains d'entre eux peuvent aussi être caractérisés par une forme un peu plus aplatie au centre (*Koch et al.*, 1981). *Corry* (1988) a répertorié à peu près 900 laccolites, principalement dans le nord des États-Unis. Leurs épaisseurs varient de quelques dizaines à quelques centaines de mètres et leurs rayons peuvent atteindre quelques kilomètres pour les plus gros (Figure 1.3 b). Ces laccolites se sont parfois mis en place les uns sur les autres formant une structure en forme d'arbre

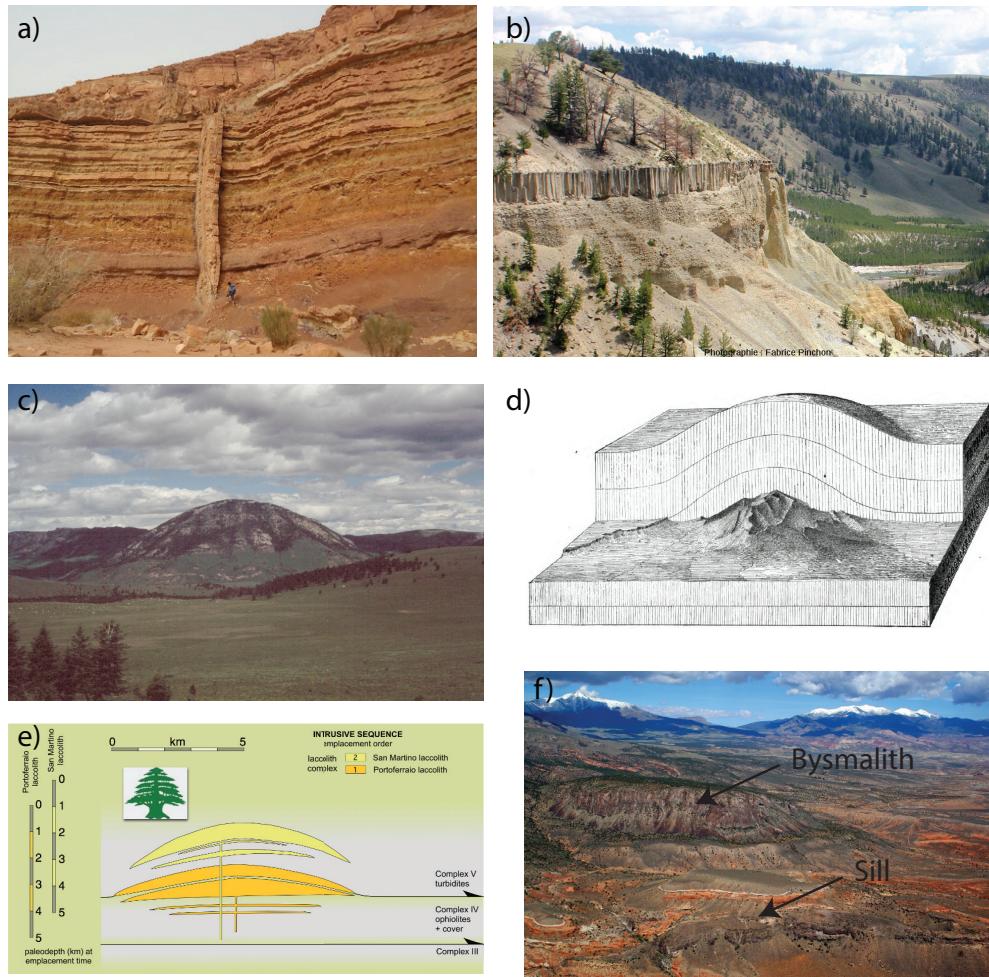


Figure 1.4: a) Dyke traversant des couches sédimentaires dans le Makhtesh Ramon, Israël; b) Sill basaltique au sein de sédiments, vallée de la Yellowstone River, parc National du Yellowstone (USA). Photographie de Fabrice Pinchon. c) Laccolite à l'érosion dans le Montana d) Schéma de l'emplacement d'un laccolite réalisé par [Gilbert \(1877\)](#). e) Schéma simplifié de la structure en arbre de noel d'un complexe de laccolite sur l'île d'Elbe, en Italie, étudiée par [Rocchi et al. \(2010\)](#). f) Intrusions à l'érosion au alentour de la montagne Hillers, dans les Henry Mountains. On peut distinguer le Black Mesa Bysmalith au centre et le Maiden Creek sill en dessous. Photographie de Jack Share

de noël (*Corry, 1988*). Cette géométrie est aussi observée sur l'île d'Elbe, en Italie, ou un complexe de neuf laccolites, exceptionnellement bien conservé, a été étudié en détail par *Rocchi et al. (2002)*.

- Les bysmalites sont d'imposants volumes cylindriques, préférentiellement composés de roche granitique, discordants (Figure 1.4 f). Ils sont notamment bordés par d'importantes failles presque verticales et peuvent atteindre quelques centaines de mètres d'épaisseur (*Johnson and Pollard, 1973*). Un exemple typique de ce type d'intrusion souvent cité est le Black Mesa Bysmalite dans les Henry Mountains (200 m d'épaisseur et 1 km de large (*Morgan et al., 2008*)).

À l'instar des batholites, de nombreuses observations de terrains proposent que ces intrusions de tailles moyennes se forment aussi par incrémentations successifs de petits volumes de magma (*Habert and De Saint-Blanquat, 2004; Horsman et al., 2005; Morgan et al., 2008*) (Figure 1.5). Cependant, les mêmes études montrent aussi que ces intrusions se forment nécessairement sur de petites échelles de temps, des échelles assez faibles pour pouvoir garder un corps chaud et liquide des premières étapes du processus d'intrusion à la solidification. Au niveau du bysmalite de Black Mesa par exemple (Figure 1.4 f), *Habert and De Saint-Blanquat (2004)* ont montré l'absence de discontinuités entre les différentes couches ainsi que l'absence de métamorphisme important dans l'encaissant indiquant un temps de mise en place de moins de 100 ans. L'absence de discontinuité au sein des différents laccolites sur l'île d'Elbe supporte aussi leur formation rapide, i.e. à la suite d'une seule injection où de plusieurs injections sur un temps assez court pour que les magmas des différentes injections coalescent (*Roni et al., 2014*).

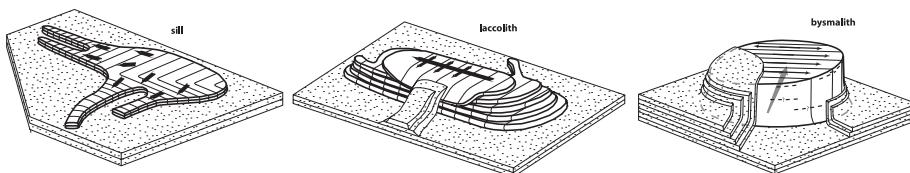


Figure 1.5: Ces diagrammes, réalisés par *Horsman et al. (2009)*, montrent la structure verticale en couche de trois intrusions à l'érosion dans les Henry Mountains. De gauche à droite: le Maiden Creek Sill (Figure 1.4 f), le Trachyte Mesa Laccolite et le Black Mesa Bysmalite (Figure 1.4 f).

1.2.2 Magmatisme intrusif sur la Lune

La lune s'est probablement formé suite à l'impact d'un corps de la taille de Mars sur la proto-Terre quelques centaines de millions d'années après la formation de la Terre, le disque de débris produit se réaccrétant ensuite en moins d'un millier d'années pour former notre satellite (*Mizutani et al., 1972; Cameron and Benz, 1991; Canup and Asphaug, 2001; Canup, 2012*). Dû aux quantités importantes d'énergie libérée durant le processus d'accrétion, on considère aujourd'hui que la Lune était partiellement fondu, sur une épaisseur encore débattue, suite à sa formation (*Elkins-Tanton et al., 2011*). Le refroidissement et la lente cristallisation fractionnée de l'océan de magma lunaire aurait ensuite conduit à la formation d'une croûte primaire par flottaison des minéraux légers de plagioclase (en particulier du pôle calcique, l'anorthite) à la surface de l'océan de magma tandis que les éléments les plus incompatibles, en particulier les éléments producteurs de chaleur, se seraient concentrés dans les derniers liquides magmatiques résiduels entre la croûte et le manteau, formé, lui, principalement de cumulats d'olivine et de pyroxène (Figure 1.6).

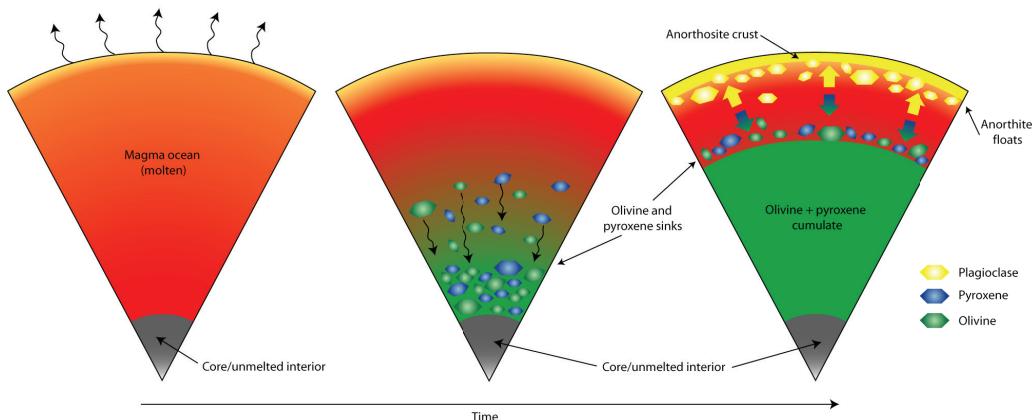


Figure 1.6: Cristallisation fractionnée de l'océan de magma et formation de la croûte primaire composé d'anorthosite. Source: LPI

Dû fait de sa composition et de la porosité résultante de 4 milliards d'années de bombardement météoritiques, la densité de la croûte lunaire est particulièrement faible (*Huang and Wieczorek, 2012; Han et al., 2014*). D'après les dernières estimations, rendues possibles grâce au champ de gravité d'une résolution sans précédente obtenu par la mission GRAIL de la NASA, la densité moyenne au niveau des terres hautes serait de 2550 kg m^{-3} (*Wieczorek et al., 2013*). Ces données ont aussi permis de réévaluer à la baisse l'épaisseur de la croûte à entre 34 et 44 km en moyenne avec une tendance à être moins

épaisse au niveau des mers lunaires.

La faible densité de sa croute et son épaisseur non négligeable ont certainement joué un rôle important sur le volcanisme lunaire. En effet, la croûte primaire formée par cristallisation de l'océan de magma étant très légère, et les magmas provenant de la fusion du manteau lunaire étant particulièrement dense, de l'ordre de 3000 kg m^{-3} (*Kiefer et al., 2012*) du à leur composition basaltique riche en oxydes métalliques, en particulier en oxyde de Fer FeO et de Titane TiO_2 , la croûte lunaire a sans aucun doute aussi été un filtre puissant à l'éruption des magmas sur la lune, leur flottabilité ne leur permettant pas d'être transporté jusqu'à la surface. *Wieczorek et al. (2001)* ont ainsi confirmé que le volcanisme à la surface est généralement lié à l'extraction d'une partie de cette croûte de faible densité, comme c'est le cas par exemple des mers lunaires qui se sont mises en place au sein de larges basins d'impacts. *Head and Wilson (1992)* ont estimé à 50 fois plus important les volumes de magma mis en place en profondeur que les volumes éruptés en surface. Cependant, bien que ce rapport puisse donner de précieuses indications sur l'évolution thermique et magmatique de la lune elle-même, il est de fait très peu contraint et la part intrusive du magmatisme lunaire est encore mal connue. La détection des déformations de surface induites par la mise en place d'intrusions magmatiques au sein de la croûte apparaît donc comme une première étape visant à la meilleure caractérisation du magmatisme intrusif lunaire.

Deux manifestations principales à la surface de la lune ont été proposées comme potentiellement résultantes de la mise en place d'intrusions magmatiques au sein de la croûte lunaire: les dômes à faible pente et les cratères à sol fracturé.

- Les dômes à faible pente sont localisés en bordure ou dans les mers lunaires, principalement sur la face visible (Figure 1.7 a, b). Une quinzaine de ces dômes, possiblement d'origine intrusive, ont été récemment décrit par *Wöhler et al. (2007)*. Bien que leur morphologie s'apparente à des laccolites terrestres, ils sont de manière générale beaucoup plus étalés que ceux sur Terre; pour une même épaisseur, l'équivalent lunaire peut ainsi être deux fois plus large que son homologue terrestre.
- Les cratères à sol fracturé sont des cratères d'impacts ayant subi des déformations suite à leur formation. À peu près 200 de ces cratères ont été répertorié par *Schultz (1976)*, principalement autour des mers lunaires (Figure 1.7 c, d, e, f). La principale caractéristique de ces cratères est leur faible profondeur par rapport à celles des cratères non déformés. En effet, certains cratères au sol fracturé peuvent être jusqu'à 2 km moins profond que leurs homologues non déformées. Leur sol, soit en forme de dôme, soit plat séparé des bords du cratère par un imposant fossé

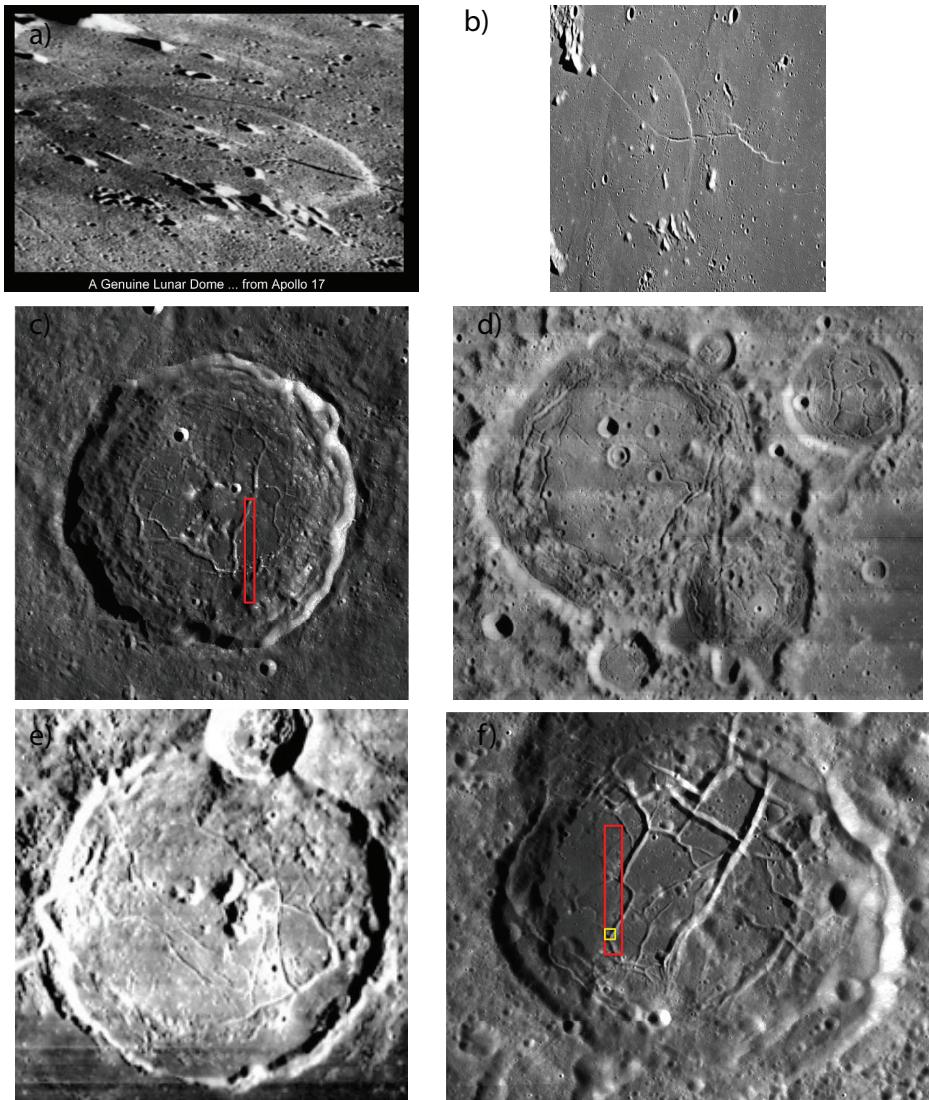


Figure 1.7: a) Dome lunaire, photo par Appolo 17 b) Apollo 15 orbital image AS15-91-12372, vue oblique du dôme Valentine. c) Cratère au sol fracturé Atlas (Classe 1). d) Cratère au sol fracturé Lavoisier (Classe 5). e) Cratère au sol fracturé Gassendi (Classe 3). f) Cratère au sol fracturé Komarov (Classe 5). Photo extraite de *Lunar Orbiter Photographic Atlas of the Moon*, NASA

circulaire, est systématiquement caractérisé par d'importants réseaux de fractures radiales, concentriques ou encore pentagonales (Figure 1.7 c, d, e, f). Basé sur leur profondeur, topographie et niveau de déformation, *Schultz* (1976) a postuler l'existence de six grandes classes de déformation. La proximité de ces cratères avec les mers lunaires, ainsi que la présence de produits volcaniques au sein de certains d'entre eux, suggère qu'ils ont été déformés suite à la mise en place de magma en profondeur sous leur sol.

1.3 Caractérisation de la mise en place d'une intrusion magmatique à faible profondeur

1.3.1 Modèle statique de déformation d'une couche élastique

Bien que la morphologie et les volumes de magma puissent être récupérés, à partir d'observations directes ou de méthodes de prospection géophysique sur Terre ou via les déformations induites à la surface des autres corps telluriques du système solaire, ces informations seules ne donnent que peu d'indications sur les mécanismes de mise en place de ces intrusions magmatiques. De nombreux travaux ont ainsi été centrés sur la modélisation des processus donnant lieu à ces déformations, dans le but de mieux comprendre le mécanisme d'intrusion d'une part, mais aussi, de déduire des observations des donnés sur le magma, les paramètres mécaniques de l'encaissant ou encore la profondeur de l'intrusion au moment de sa mise en place.

La propagation d'un dyke dans un milieu élastique a été beaucoup étudiée (*Lister and Kerr*, 1991; *Rubin*, 1995). En particulier, *Lister and Kerr* (1991) ont montré que, à l'exception de la tête du dyke où les contraintes élastiques induites par les roches encaissantes jouent un rôle important, la dynamique du magma au sein du dyke est contrôlée par un équilibre entre la flottabilité et les pertes de charges associées aux frottements visqueux sur les parois du conduit. On a vu qu'un dyke peut se transformer en sill si celui-ci rencontre sa zone de flottabilité neutre; bien que la dynamique des dykes et des sills soit comparable à forte profondeur (*Lister and Kerr*, 1991; *Cruden et al.*, 2012), à faible profondeur, la forme des laccolites suppose que les intrusions magmatiques se mettent en place principalement par flexion des couches sus-jacentes (*Johnson and Pollard*, 1973). Un modèle, populaire en science planétaire, consiste à modéliser ces laccolites par la déformation d'une plaque mince élastique, de longueur fixée et égale à la taille de l'intrusion, soumise à une pression donnée (*Pollard and Johnson*, 1973). Dans ces modèles statiques, cette pression, don-

nant lieu à la déformation, est soit prise constante sur la taille de l'intrusion et égale au poids du magma (*Pollard and Johnson*, 1973; *Wichman and Schultz*, 1996; *Jozwiak et al.*, 2012), soit imposé suivant un profil décrivant la perte de charge associée à un écoulement visqueux (*Kerr and Pollard*, 1998; *Wöhler et al.*, 2009). Cependant, dans aucun des cas, cette pression n'est reliée aux paramètres de l'écoulement lui-même, i.e. volume ou taux d'injection. De plus, ces modèles ne fournissent pas un cadre théorique suffisant à la compréhension de la dynamique de l'intrusion et sont donc incapables d'expliquer la diversité des formes et des tailles observées. Enfin, ils considèrent la flexion de la couche sus-jacente comme unique pression motrice à l'écoulement, sans considérer le poids du magma lui-même, qui doit pourtant nécessairement jouer un rôle sur la mise en place de l'intrusion.

1.3.2 Inférence sur la dynamique à partir de leur géométrie

En l'absence d'un modèle dynamique, la géométrie des intrusions répertoriées a souvent été utilisée pour en déduire des indications sur les processus de mise en place et de croissance de ces intrusions. Ainsi, en utilisant les données répertoriées sur les laccolites par *Corry* (1988), *McCaffrey and Petford* (1997) propose une loi de puissance empirique pour l'épaisseur des intrusions h_0 en fonction de leur longueur R , $h_0 = bR^a$ où a est l'exposant de la loi de puissance et b une constante. Un exposant supérieur à l'unité indique que l'intrusion croît préférentiellement en s'épaississant tandis qu'un exposant inférieur à l'unité indique qu'elle croît plutôt par étalement.

Les laccolites répertoriées par *Corry* (1988) montrent un exposant $a < 1$ (0.88 ± 0.1), interprété comme reflétant l'étalement de l'intrusion sur une certaine distance sous forme d'un sill avant son épaisseissement (Figure 1.8). Ce modèle est cohérent avec le modèle en deux étapes couramment accepté pour la mise en place des laccolites (*Johnson and Pollard*, 1973; *McCaffrey and Petford*, 1997). Premièrement, le magma s'étale latéralement au niveau de sa zone de flottabilité neutre, i.e. $a < 1$ jusqu'à ce qu'un sill, caractérisé par un rapport d'aspect assez large, soit formé. Ensuite, lorsque le sill est assez large, il s'épaissit par flexion des couches sus-jacentes pour former un laccolite caractérisé par une valeur de $a > 1$ (*Johnson and Pollard*, 1973; *Koch et al.*, 1981). Si la roche sus-jacente est soumise à des contraintes trop importantes, des failles se forment au niveau des bords du sill et celui-ci s'épaissit uniformément sur toute sa surface formant un bysmalite (*Corry*, 1988). Dans la continuité de l'étude de *McCaffrey and Petford* (1997), *Rocchi et al.* (2002) ont réalisé une étude détaillée du complexe intrusif de l'île d'Elbe en Italie et ont trouvé un exposant a supérieur à l'unité, i.e. ~ 1.5 , interprété comme

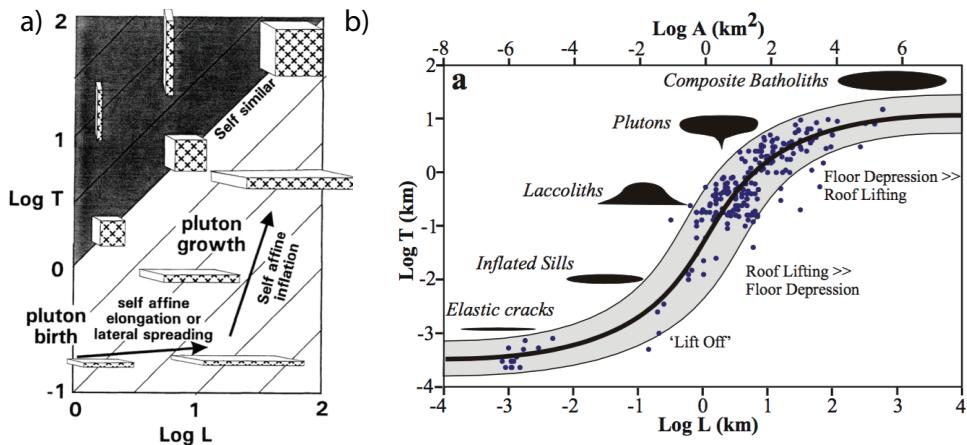


Figure 1.8: a) Schéma de la formation des laccolites en deux étapes par McCaffrey and Petford (1997). Épaisseurs en fonction de leur longueur de différents types d'intrusions magmatiques à différentes locations. Figure extraite de Cruden et al. (2012).

étant la preuve de l'existence d'une phase dominée par l'épaississement de l'intrusion dans la croissance de ces laccolites.

Des modèles plus récents conçoivent plutôt la formation des laccolites par empilements successifs de sills, de grand rapport d'aspect, plutôt que par l'injection d'un seul volume de magma fini à un temps donné (Menand, 2011). En effet, ces modèles sont étayés par les expériences de Kavanagh et al. (2006) (Section 1.1.3) où il est montré qu'un sill peut se mettre en place à l'interface entre deux couches de rigidité différentes, la rigidité de la couche sus-jacente étant plus importante que celle de la couche sous-jacente. Dès lors, la mise en place d'un sill, en refroidissant, procure un environnement favorable à la mise en place d'un nouveau sill, soit au-dessus si la rigidité du sill solidifié est inférieure à celle de la roche sus-jacente, soit en dessous dans le cas contraire. Ce modèle de croissance a aussi été suggéré par de récentes études structurales et stratigraphiques, notamment au niveau des intrusions de tailles intermédiaires dans les Henry Mountains (Horsman et al., 2005; Morgan et al., 2008; Horsman et al., 2009; Menand, 2011). Ce modèle, à la différence des modèles statiques exposés plus haut, a aussi l'avantage de pouvoir expliquer la structure aplatie au niveau du centre de certains laccolites (Morgan et al., 2008). Cependant, ce modèle ne fournit pas de mécanisme ni ne permet d'expliquer l'origine de la loi de puissance caractéristique de la géométrie de ces intrusions. De plus, il ne permet pas non plus de relier la géométrie finale de l'intrusion aux propriétés physiques de l'écoulement (volume, taux d'injection).

Cruden and McCaffrey (2002) ont réuni des données sur une plus grande plage de longueurs, de petits filons de quelques dizaines de mètres à des batholites de quelques centaines de kilomètres (Figure 1.8) et proposent que l'épaisseur en fonction de la longueur des intrusions magmatiques forme une distribution en forme de sigmoïde (dans une échelle logarithmique), avec une pente maximum de 1.5 caractéristiques des laccolites. Cependant, aucune théorie sous-jacente ne soutient cette observation. De plus, les données de *Cruden et al.* (2012) sur les larges sills mafiques contredisent cette affirmation (Figure 1.3).

1.3.3 Discussion

Bien que de nombreux modèles ont été proposés pour essayer de rendre compte des observations, peu d'entre eux s'intéressent à la dynamique même de l'intrusion qui permettrait cependant de relier la morphologie de ces intrusions aux propriétés physiques de l'écoulement (volume ou taux d'injection). Afin de comprendre la morphologie des intrusions peu profondes, il apparaissait ainsi nécessaire de s'intéresser à la dynamique d'un tel écoulement.

Michaut (2011) a donc développé un modèle théorique d'étalement d'un magma visqueux sous une couche élastique d'épaisseur constante continuelllement nourrie par un conduit vertical en son centre. Ce modèle diffère de ses prédecesseurs par sa capacité à traiter la dynamique même de l'intrusion ainsi que le poids du magma comme un moteur de l'écoulement. Les résultats et la capacité de ce modèle à reproduire les observations sont discutés dans le chapitre suivant.

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CHAPTER 2

Isoviscous elastic-plated gravity current model for shallow magmatic intrusion

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Michaut (2011) proposed a new model for the spreading of a shallow depth intermediate-size intrusion, where magma is continuously injected at the center and is accommodated by the bending of the overlying strata. In particular, the model differs from previous ones by considering both the dynamics of the emplacement itself, in a sense that the radius is self-consistently determined, and the driving force associated with the magma weight. Both were neglected in older models. In the original paper from *Michaut* (2011), the model was

derived in both cartesian and axisymmetric geometry and the results were presented in 2D. A similar model in 2D with an additional fracture criterion at the tip of the intrusion has been derived by *Bunger and Cruden* (2011) and *Hewitt et al.* (2014) discussed more precisely the dynamics at the contact line and the case of an elastic-plated gravity current spreading over an inclined plane. In this chapter, we present a summary of the model and the results for the spreading of an isoviscous elastic-plated gravity current over a rigid horizontal surface in an axisymmetrical geometry. Results in this geometry have been thoroughly studied by *Lister et al.* (2013) and this model will constitute the reference for more elaborate models in the manuscript.

2.1 Theoretical model

The model considers an isoviscous elastic-plated gravity current, i.e. an isoviscous fluid of viscosity η_h and density ρ_m spreading beneath a thin elastic sheet of thickness d_c and above a semi infinite rigid layer (*Michaut*, 2011; *Bunger and Cruden*, 2011) (Figure 2.1). The fluid is injected continuously at the base and center of the current at a rate Q_0 through a cylindrical conduit of diameter a .

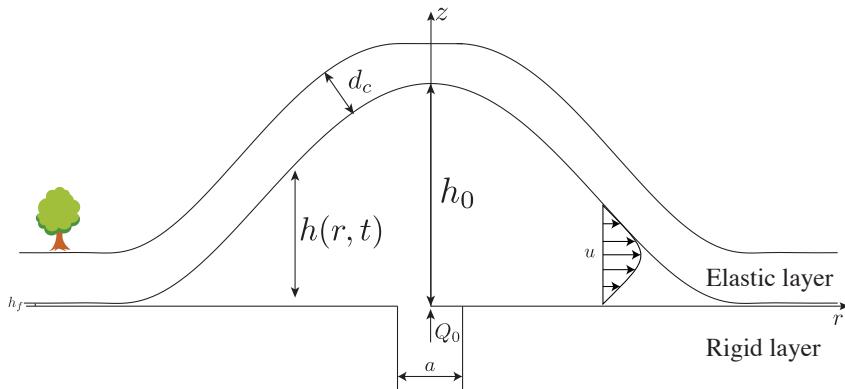


Figure 2.1: Model geometry and parameters.

2.1.1 Governing equation

Driving pressure

The intrusion develops over a length scale Λ that is much larger than its thickness H ($\varepsilon = H/\Lambda \ll 1$). In the laminar regime and in axisymmetrical

coordinates (r, z) , the Navier-Stokes equations within the lubrication assumption are

$$-\frac{\partial P}{\partial r} + \frac{\partial}{\partial z} \left(\eta_b \frac{\partial u}{\partial z} \right) = 0 \quad (2.1)$$

$$-\frac{\partial P}{\partial z} - \rho_m g = 0 \quad (2.2)$$

where $u(r, z, t)$ is the radial velocity, g is the standard acceleration due to gravity and $P(r, z, t)$ is the pressure within the fluid. Integration of (2.2) thus gives the total pressure $P(r, z, t)$ within the flow. When the vertical deflection $h(r, t)$ of the upper elastic layer is small compared to its thickness d_c , i.e $h \ll d_c$, we can neglect stretching of the upper layer and only consider bending stresses. Therefore, the total pressure $P(r, z, t)$ at a level z in the intrusion is the sum of four contributions: the weight of the magma and of the upper layer, the bending pressure P_b and the atmospheric pressure P_0

$$P = \rho_m g(h - z) + \rho_r g d_c + P_b + P_0 \quad (2.3)$$

where $h(r, t)$ is the intrusion thickness and ρ_r the density of the surrounding rocks. The bending pressure is given by the force per unit area that is necessary for a vertical displacement h of the thin elastic plate (*Turcotte and Schubert, 1982*)

$$P_d = D \nabla_r^4 h \quad (2.4)$$

where D is the flexural rigidity of the thin elastic layer, that depends on the Young's modulus E , Poisson's ratio ν^* and on the elastic layer thickness d_c as $D = E d_c^3 / (12(1 - \nu^{*2}))$.

Velocity field

At the contact with the elastic sheet $z = h(r, t)$, the no-slip boundary condition hold and then, the tangential velocity is zero and the normal velocity is the change in height ($\partial h / \partial t$). With \vec{n} the normal to the surface and \vec{t} the tangent, we have

$$\vec{n} \cdot (u, w) = \frac{\partial h}{\partial t} \quad (2.5)$$

$$\vec{t} \cdot (u, w) = 0. \quad (2.6)$$

The tangent vector is $\vec{t} = (1, \partial h / \partial r)$. However, within the lubrication assumption, the vertical component of the tangent vector scales as ε and thus, is negligible compared to the radial component. Therefore, the boundary condition (2.6) reduces to $u(r, z = h, t) = 0$. At the base of the flow, the same boundary condition hold and $u(r, z = 0, t) = 0$.

Equation (2.1) is integrated twice as a function of z using these boundary conditions and the horizontal velocity is

$$u(r, z, t) = \frac{1}{2\eta_h} \frac{\partial P}{\partial r} (z^2 - hz) \quad (2.7)$$

Injection rate

The effective overpressure ΔP^* driving the flow in the feeder conduit decreases as the intrusion thickens and is given by

$$\Delta P^* = \Delta P - \rho_m g h_0 \quad (2.8)$$

where $h_0(t)$ is the maximum intrusion thickness at the center $r = 0$ and ΔP is the initial driving pressure or the overpressure at the base of the dyke ($z = -Z_c$).

In (2.8), the bending pressure at the center, which scales as $Dh_0(t)/R(t)^4$ where $R(t)$ is the current radius, has been neglected. Although it tends to infinity at the initiation of the flow, it rapidly vanishes as the current spreads and the hydrostatic pressure $\rho_m g h_0$ becomes the main contribution to the pressure at the center. In addition, the model assumes a large aspect ratio for the flow and does not consider the initiation of the flow.

Finally, assuming a Poiseuille flow within the cylindrical feeding conduit, the vertical injection velocity $w_i(r, t)$ and injection rate $Q(t)$ are given by

$$w_i = \begin{cases} \frac{\Delta P^*}{4\eta_h Z_c} \left(\frac{a^2}{4} - r^2 \right) & r \leq \frac{a}{2} \\ 0 & r > \frac{a}{2} \end{cases} \quad (2.9)$$

$$Q = Q_0 \left(1 - \frac{\rho_m g h_0}{\Delta P} \right) \quad (2.10)$$

where $Q_0 = (\pi \Delta P^* a^4) / (128 \eta_h Z_c)$.

Mass conservation

The fluid is assumed incompressible and a global statement of mass conservation gives

$$\frac{\partial h}{\partial t} + \frac{1}{r} \frac{\partial}{\partial r} \left(r \int_0^h u dz \right) = w_i \quad (2.11)$$

and using (2.7), we find that the equation for the evolution of the thickness in time and space reads

$$\frac{\partial h}{\partial t} = \frac{\rho_m g}{12\eta_h r} \frac{\partial}{\partial r} \left(r h^3 \frac{\partial h}{\partial r} \right) + \frac{D}{12\eta_h r} \left(r h^3 \frac{\partial}{\partial r} \nabla_r^4 h \right) + w_i. \quad (2.12)$$

It is composed of three different terms on the right hand side. The first term represents gravitational spreading, i.e. spreading of the current under its own weight. The second term represents the squeezing of the flow by the upper elastic layer. Both term are negative and induces spreading. The last term represents fluid injection and is positive.

2.1.2 Dimensionless equations

Equation (2.12) is nondimensionalized using a horizontal scale Λ , a vertical scale H and a time scale τ given by

$$\Lambda = \left(\frac{D}{\rho_m g} \right)^{1/4} \quad (2.13)$$

$$H = \left(\frac{12\eta_h Q_0}{\rho_m g \pi} \right)^{1/4} \quad (2.14)$$

$$\tau = \frac{\pi \Lambda^2 H}{Q_0} \quad (2.15)$$

where scales are chosen such that $Q_0 = \pi \Lambda^2 H / \tau$. The length scale represents the flexural wavelength of the upper elastic layer, i.e. the length scale at which bending stresses and gravity contribute equally to flow. The height scale H is the thickness of a typical gravity current and the time scale τ is the characteristic time to fill up a cylindrical flow of radius Λ and thickness H at constant rate Q_0 . In addition, we can define a horizontal velocity scale $U = \Lambda / \tau = (\rho_m g H^3) / (12\eta_h \Lambda)$.

The dimensionless equation is

$$\begin{aligned} \frac{\partial h}{\partial t} &= \frac{1}{r} \frac{\partial}{\partial r} \left(r h^3 \frac{\partial h}{\partial r} \right) + \frac{1}{r} \left(r h^3 \frac{\partial}{\partial r} \nabla_r^4 h \right) \\ &+ \frac{32}{\gamma^2} \left(\frac{1}{4} - \frac{r^2}{\gamma^2} \right) \left(1 - \frac{h_0}{\sigma} \right) \end{aligned} \quad (2.16)$$

where the last term is replaced by zero for $r > \gamma/2$. γ and σ are two dimensionless numbers that control the dynamics of the flow

$$\gamma = \frac{a}{\Lambda} \quad (2.17)$$

$$\sigma = \frac{\Delta P}{\rho_m g H}. \quad (2.18)$$

γ is the dimensionless radius of the conduit, it does not significantly influence the flow and is set to 0.02 in the following (*Michaut and Bercovici, 2009; Michaut, 2011*). σ is the normalized pressure head, i.e., the ratio between the initial overpressure driving the flow and the weight of the magma at the center.

2.1.3 Need for regularization

One of the main mathematical difficulty in solving equation (2.16) arises at the contact line. Indeed, the assumption that the thickness of the fluid tends to zero at the contact line leads to divergent viscous stresses, i.e. $\eta_h \partial u / \partial z \rightarrow \infty$ and hence, the theoretical immobility of the blister (*Flitton and King*, 2004; *Lister et al.*, 2013; *Hewitt et al.*, 2014). This problem, known as the contact-line paradox, is a well known problem for surface-tension driven flow such as the spreading of a water droplet (*Bertozzi*, 1998; *Snoeijer and Andreotti*, 2013).

The formal proof have been derived by *Flitton and King* (2004) and can be derived as follow. Suppose that (2.16) has a solution and the solution has the form $h \sim A(t)(R(t) - r)^\alpha$ near the contact line. As $r \rightarrow R(r)$, the bending term dominates the gravitational term and (2.16) reduces to

$$\frac{\partial h}{\partial t} = \frac{1}{r} \frac{\partial}{\partial r} \left(r h^3 \frac{\partial}{\partial r} \nabla_r^4 h \right). \quad (2.19)$$

Injecting the solution into (2.19) and keeping only the leading powers of $R - r$ gives

$$\begin{aligned} \frac{\partial R}{\partial t} A \alpha (R - r)^{\alpha-1} + \frac{\partial A}{\partial t} (R - r)^\alpha &= A^4 \alpha (\alpha - 1) (\alpha - 2) \\ &\quad (\alpha - 3) (\alpha - 4) (\alpha - 5) (R - r)^{4\alpha-6} \end{aligned}$$

The time derivative is locally dominated by its convective part at the tip, the second term on the left is small compared to the first and therefore, by equating the exponent of $R - r$, we obtain $\alpha = 5/3$ and then

$$\frac{\partial R}{\partial r} = -\frac{280}{243} A^3. \quad (2.20)$$

It shows that (2.16) can only have solutions with retreating contact line ($dR/dt < 0$) but not with advancing contact line ($dR/dt > 0$) (*Lister et al.*, 2013; *Flitton and King*, 2004).

To mitigate this problem, one common approach is to add a thin prewetting film, with thickness h_f such that $h \rightarrow h_f$ as $r \rightarrow \infty$. While the solution will depend upon the prewetting film thickness h_f and will not show any convergence properties when $h_f \rightarrow 0$, we will see that the dependence in h_f is weak and the difference between different values for h_f will be relatively small (*Lister et al.*, 2013; *Hewitt et al.*, 2014). Unless otherwise specified, we will consider $h_f = 5 \cdot 10^{-3}$ in the manuscript.

2.2 Results

For a small prewetting film thickness, i.e. $h_f \ll 1$, the numerical resolution of the equation (2.16) shows three asymptotic spreading regimes: a bending

regime where gravity is negligible, a viscous gravity current regime where bending is negligible and a regime of lateral propagation (*Michaut, 2011; Bunger and Cruden, 2011; Lister et al., 2013*). In the following, we present the shape of the flow as well as scaling laws that predict the evolution of the thickness at the center $h_0(t)$ and the radius $R(t)$ in each regime.

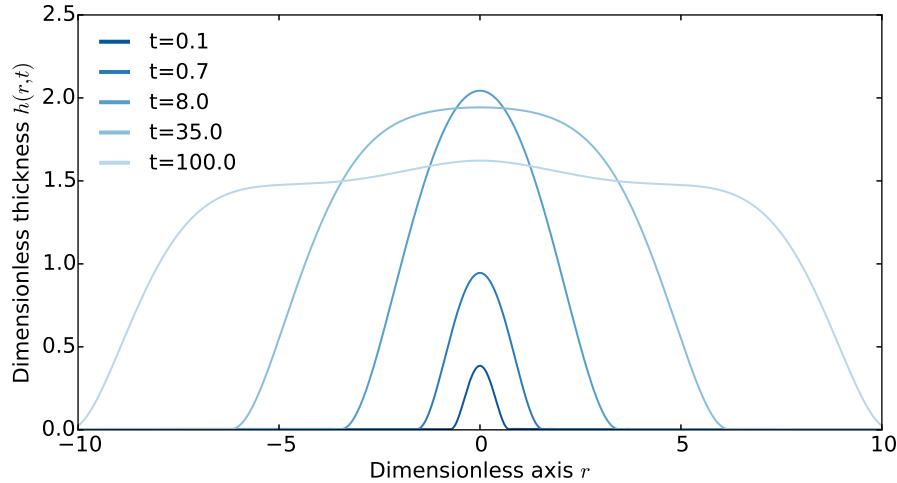


Figure 2.2: Shape of the flow, i.e. thickness $h(r, t)$ as a function of the radial axis r at five different times indicated on the plot. Variables are dimensionless and one needs to multiply by the characteristic scales (thickness, length or time given by (3.23), (3.22) or (3.24)) to obtain dimensional values. For $t < 10$, the intrusion is in the bending regime whereas for $t > 10$ the intrusion is in the gravity current regime.

2.2.1 Bending regime

At early times, when $R \ll \Lambda$, gravity is negligible and the dynamics of the spreading is governed by the bending of the upper layer. In addition, if $h_0 \ll \sigma$, the overpressure ΔP driving the flow is much larger than the weight of the blister at the center and the injection rate can be considered constant.

In that case, the spreading is very slow and the interior has uniform dimensionless pressure $P = \nabla_r^4 h$. The flow is bell-shaped and its thickness is given by

$$h(r, t) = h_0(t) \left(1 - \frac{r^2}{R^2(t)}\right)^2 \quad (2.21)$$

with $h_0(t)$ the thickness of the intrusion at the center (*Michaut, 2011; Lister et al., 2013*). In this regime, *Lister et al.* (2013) have shown that the spreading

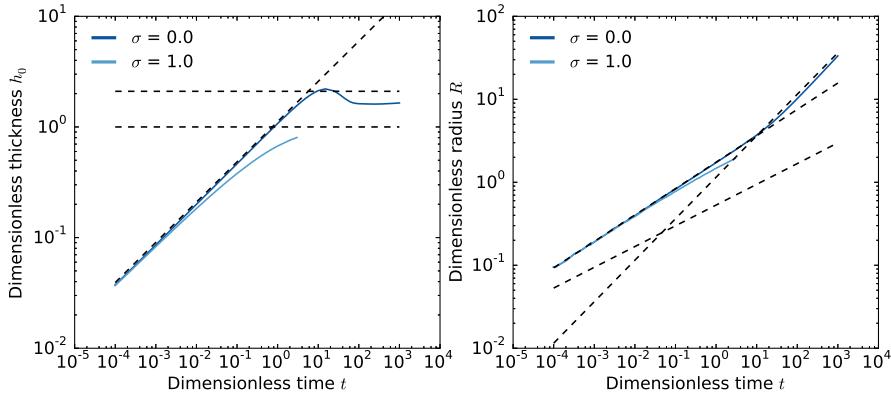


Figure 2.3: Left: Dimensionless thickness at the center h_0 versus dimensionless time t for different dimensionless number σ indicated on the plot. Dashed-lines represent the scaling laws in the different regimes. Right: Dimensionless radius R versus dimensionless time t for the same dimensionless number σ . Dashed-lines represent the scaling laws in the different regimes.

is controlled by the propagation of a peeling by bending wave at the intrusion front with dimensionless velocity c

$$c = \frac{\partial R}{\partial t} = h_f^{1/2} \left(\frac{\kappa}{1.35} \right)^{5/2} \quad (2.22)$$

where $\kappa = \partial^2 h / \partial r^2$ is the dimensionless curvature of the interior solution. Using the propagation law (3.50) and the form of the interior solution (3.49), they find that the radius and the height of the intrusion evolve following

$$R(t) = 2.2 h_f^{1/22} t^{7/22} \quad (2.23)$$

$$h_0(t) = 0.7 h_f^{-1/11} t^{8/22} \quad (2.24)$$

where the numerical pre-factors have been matched to our simulations. The bell-shaped morphology of the flow in this regime is very close to the dome-shaped morphology of solidified laccoliths (*Michaut, 2011*).

2.2.2 Gravity current regime

In contrast, when the radius R becomes much larger than Λ ($R \gg \Lambda$), the weight of the intrusion becomes dominant over the bending terms. The dimensionless pressure is given by the hydrostatic pressure $P = h$ and the intrusion enters a classical viscous gravity current regime where bending terms only affect the solution near the intrusion edge (*Huppert, 1982; Michaut, 2011*;

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Lister et al., 2013). In this second regime, while the thickness tends to a constant, the radius evolves as $t^{1/2}$. The flow is therefore characterized by a small aspect-ratio and a constant thickness disk-like morphology close to the one shown by large mafic sills.

In between the bending and gravity regime, *Lister et al.* (2013) also describe a short intermediate regime where the peeling by bending continues to control the propagation but where, due to the increasing effect of gravity, the flow shows an interior flat-topped region. This flat-topped morphology is also observed for many laccoliths (*Koch et al.*, 1981; *Bunger and Cruden*, 2011).

2.2.3 Lateral propagation

Once $h_0 \rightarrow \sigma$, the flow is thick enough to compensate for the initial overpressure. The thickness at the center remains constant and the flow enters a regime of lateral propagation, where only its radius $R(t)$ is to increase (*Michaut*, 2011). In this regime, except at the center when it redistributes the pressure over a length scale Λ , the bending term is negligible compared to the gravitational term. *Michaut* (2011) has shown that in this regime, the thickness is constant and the radius evolves as $t^{1/4}$

$$R(t) = \left(\frac{\sigma^3 t}{4\pi} \right)^{1/4} \quad (2.25)$$

$$h_0 = \sigma \quad (2.26)$$

2.3 Application to the spreading of shallow magmatic intrusions

2.3.1 Observations versus predictions on the Earth

Dataset

Corry (1988) has made an extensive catalog of 900 laccoliths across the world. In particular, *Corry* (1988) provides for the thickness and the radius of 168 laccoliths among which, 40 are also given with an estimation of the intrusion depth. These laccoliths, who are mainly felsic in composition, show thicknesses that range from 100 meters to 10 km with radii in between 1 and 10 km. While most of the data are located in the United States ($\sim 90\%$), the different laccoliths are widely spread among the territory and variations in the parameters between different laccoliths are most likely to be important.

Chapter 2. Isoviscous elastic-plated gravity current model for shallow magmatic intrusion

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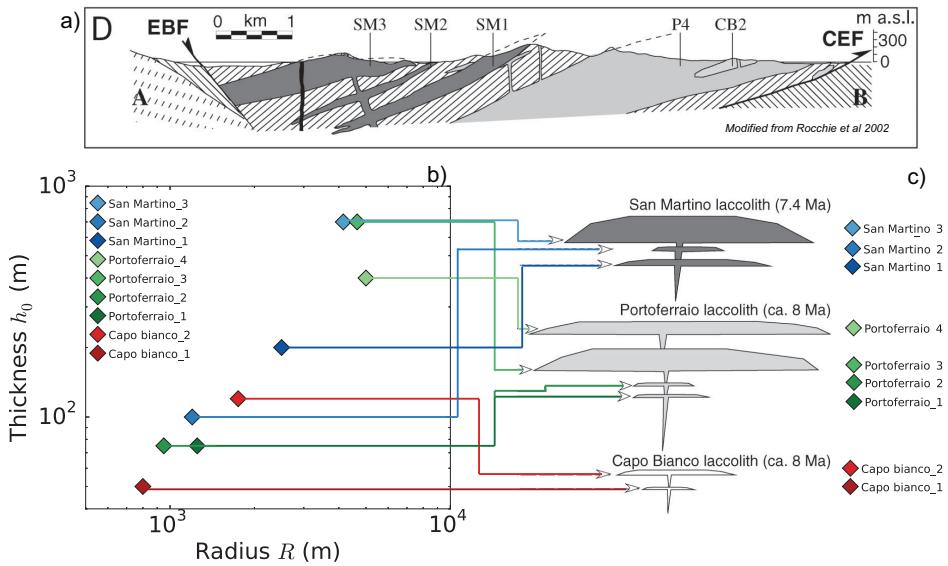


Figure 2.4: a): Cross section of western and central Elba Island where we can see the christmas tree structure of the laccolith complex and the main laccolith units visible at the surface. b) Thickness versus radius of the different laccolith units. c) Sketch of the corresponding location of these laccoliths within the christmas tree structure shortly after their formations. Figure modified from *Rocchi et al. (2002)*.

Therefore, in addition to the data from *Corry (1988)*, we also consider in this study the data provided by *Rocchi et al. (2002)* on 9 laccoliths nested in a christmas tree structure at Elba Island, Italy (Figure 2.4). The detailed mapping and reconstruction of tectonic history made by *Rocchi et al. (2002)* provides for the parameters of each intrusive layer in the laccolith complex. In addition, for this dataset, each laccolith is part of a larger intrusive system, and hence variability of the model parameters should be limited, except for the overlying elastic layer thickness, taken to be the intrusion depth, whose variation between laccoliths is given by *Rocchi et al. (2002)*. The dispersion is much smaller for this dataset; the radius ranges from 1 to 10 km and the thickness from 40 m to 1 km.

Finally, we also account for 25 large mafic sills whose thicknesses and radii are given by *Cruden et al. (2012)*. In order to account for the intrinsic scale of different settings for each intrusion and compare them to the model, the data have first to be nondimensionalized using characteristic values for each intrusion parameters and also their depth, when absent from the catalog.

Range of values for the parameters

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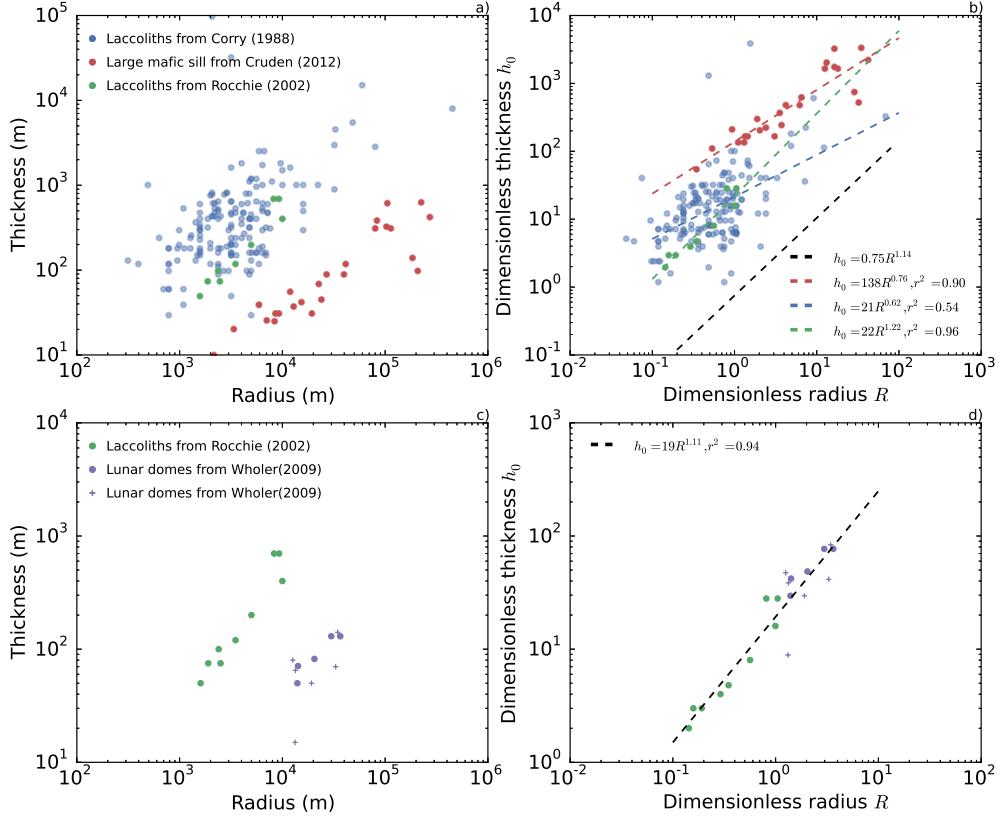


Figure 2.5: a) Thickness at the center h_0 (m) versus the radius R (m) for magmatic intrusions from different datasets indicated on the plot. b) Dimensionless thickness as a function of dimensionless radius, characteristic thickness, and length are calculated from (3.23) and (3.22). Dashed lines: predicted scaling law from the simulations (black) and best fit for the power law $h_0 = aR^b$ for each dataset obtained from a linear least-square regression in log-log space. r^2 is the squared of the correlation coefficient, i.e. Pearson product-moment correlation coefficient which gives an indication on the goodness of the fit; 1 is total positive correlation and 0 is no correlation. We use $\rho_m = 2500 \text{ kg m}^{-3}$, $Q_0 = 2.5 \cdot 10^3 \text{ m}^3 \text{ s}^{-1}$ and $\eta_h = 10^6 \text{ Pa s}$ for felsic laccoliths and $\rho_m = 2900 \text{ kg m}^{-3}$, $Q_0 = 0.1 \text{ m}^3 \text{ s}^{-1}$ and $\eta_h = 10^2 \text{ Pa s}$ for their mafic counterparts. Unless the intrusion depth is given by the dataset, we use $d_c = 1500 \text{ m}$ and $g = 9.81 \text{ m s}^{-2}$. c) and d), same plots but where we compared the laccoliths from [Rocchi et al. \(2002\)](#) to a set of low-slope lunar domes given by [Wöhler et al. \(2009\)](#). Lunar domes are nondimensionalized using $g = 1.62 \text{ m s}^{-2}$, $\rho_m = 3000 \text{ kg m}^{-3}$, $Q_0 = 1 \text{ m}^3 \text{ s}^{-1}$, $\eta_h = 1 \text{ Pa s}$ and d_c , which is not given in the dataset, is set to 1500 m. Purple dots correspond to morphometry reevaluated with the LOLA instrument topography and crosses to the original data. In all cases, the Poisson's ratio is $\nu^* = 0.25$.

In terrestrial settings, magma density ρ_m depends mainly on its composition and varies between 2500 kg m^{-3} for felsic lavas to 3000 kg m^{-3} for more mafic lavas. Reported intrusion depths, which is set to 1.5 km otherwise, vary from 180 to 2200 m for laccoliths in *Corry* (1988) and from 1.9 to 3.7 km for laccoliths at Elba Island. Hence, for a Young's modulus value of 10 GPa, the characteristic length scale Λ varies between $\sim 1 \text{ km}$ and $\sim 7 \text{ km}$ for laccoliths. The density does not affect much the value of Λ and the characteristic length scale for large mafic sills, whose depths are not reported in *Cruden et al.* (2012) and set to 1.5 km, is equal to $\sim 3 \text{ km}$.

On Earth, laccoliths are generally formed by relatively evolved lavas that may have differentiated from primitive magma in deep crustal magma chambers, located some 5 to 15 km below the surface. The overpressures driving magma ascent are typically 20 to 50 MPa (*Stasiuk et al.*, 1993; *Barmin et al.*, 2002), which gives overpressure gradients of $\sim 10^3 \text{ Pa m}^{-1}$. Lava viscosity at eruption temperature η_h depends mainly on its composition and water content; close to its liquidus temperature, it can varies from 10^2 Pa s for mafic lavas to 10^6 Pa s for felsic lavas (*Shaw*, 1972; *Giordano et al.*, 2008; *Whittington et al.*, 2009; *Chevrel et al.*, 2013). *Wada* (2007) show that the dyke width tends to increase with viscosity to the power 1/4 (*Kerr and Lister*, 1995); mafic lavas with viscosity 10^2 Pa s at eruption temperature tend to form dyke 1 m wide, while felsic magmas, with viscosities of $10^6\text{-}10^7 \text{ Pa s}$ at eruption temperature, tend to form dykes 100 m wide. For the same overpressure gradient, plugin in these parameters in Q_0 (2.10) gives an injection rate close to $0.1 \text{ m}^3 \text{ s}^{-1}$ for mafic lavas and $2.5 \cdot 10^3 \text{ m}^3 \text{ s}^{-1}$ for felsic lavas. The height scale H is thus ~ 25 for felsic laccoliths and $\sim 0.1 \text{ m}$ for large mafic sills.

The model also considers a thin pre-wetted film of thickness h_f whose meaning in the application to the spreading of laccolith is unclear. In particular, the model shows no convergence when h_f tends to zero (*Lister et al.*, 2013) and therefore, the thickness h_f might be linked to some structural length scale at the front of the laccolith or to the natural imperfection of the flow geometry. For the purpose of the application, we choose a film thickness of 1 mm, i.e. the minimum length scale with physical significance for the spreading of laccoliths which give a dimensionless h_f that varies between 10^{-2} and 10^{-4} . In the following, we set h_f to 10^{-3} .

Dimensionless data and comparison with the model

Each magmatic intrusion unit is made dimensionless using its characteristic length scale Λ , which depends upon the intrusion depth, and its characteristic height scale, which is either $H = 25 \text{ m}$ for felsic laccoliths or $H = 0.1 \text{ m}$ for large mafic sills (Figure 2.5). First, the dimensionless radius of laccoliths

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Table 2.1: Range of values for the model parameters

Parameters	Symbol	Earth	Moon	Unit
Depth of intrusion	d_c	0.2 – 2.7	1.5	km
Young’s Modulus	E	10	10	GPa
Poisson’s ratio	ν^*	0.25	0.25	
Gravity	g	9.81	1.62	m s^{-2}
Magma density	ρ_m	2500 – 2900	2900	kg m^{-3}
Magma viscosity	η_h	$10^2 – 10^6$	1	Pa s
Feeder dyke width	a	1 – 100	10	m
Depth of the melt source	Z_c	1 – 10	500	km
Initial overpressure	ΔP	20 – 50	50	MPa
Injection rate	Q_0	$0.1 – 10^3$	10^4	$\text{m}^3 \text{s}^{-1}$
Characteristic scales	Symbol	Earth	Moon	Unit
Height scale	H	0.1 – 25	2	m
Length scale	Λ	1 – 7	3.3	km
Time scale	τ	$10^{-1} – 10$	$10^{-2} – 1$	years

at Elba Island and 95% of those from [Corry \(1988\)](#) are smaller than 4 consistent with their arrest in the bending regime. The prediction of the model for the evolution of the thickness h_0 of the current as a function of its radius R can be easily derived from the scaling laws (3.51) and (3.52) and should follow

$$h_0 \sim 0.3 h_f^{-1/7} R^{8/7} \quad (2.27)$$

in agreement with the power law relationship $h_0 = bR^a$ proposed initially by [McCaffrey and Petford \(1997\)](#) (Section 1.3.2). To characterize the mean trend in each population, we make use of a linear least-square regression in log-log space to obtain a value for the coefficient a and b that best fit the observations. We found $h_0 = 21R^{1.22}$ for the laccoliths at Elba island which is very close to $R^{1.14}$ predicted by the model (Figure 2.5). Actually, the geometry of these laccoliths is not well known and probably not perfectly axisymmetric. [Hewitt et al. \(2014\)](#) found that for a two dimensional flow, $h_0 \propto h_f^{-1/7} L^{10/7}$ where L is the half-length of the flow ($10/7 \sim 1.43$). The best fit value for the coefficient a then nicely inserts between the expected values for the two geometries as noted by [Michaut \(2011\)](#). In contrast, the prediction for the coefficient b is much smaller than the predicted value. Even for $h_f = 10^{-2}$, which would be an upper bound for this parameter, the model predict $b = 0.15$, which is three

orders of magnitudes smaller than the observation (Figure 2.5). Matching the data to the model will require using a viscosity η_h for the magma abnormally high, i.e. $\eta_h \sim 10^{12}$ Pa s or unreasonable injection rate, i.e. $Q_0 \sim 1 \text{ km}^3 \text{ s}^{-1}$.

The best fit power law relationship for the laccoliths from [Corry \(1988\)](#) is $h_0 = 21R^{0.62}$ (Figure 2.5, $r^2 = 0.54$). In that case, the exponent a is smaller than one and does not agree with the model. This value for a , slightly smaller than the value calculated directly on the data by [McCaffrey and Petford \(1997\)](#), was interpreted as reflecting the two-stage growth process historically invoked for the formation of laccolith (Section 1.3.2). However, the dispersion in the data is much important than in the observation from [Rocchi et al. \(2010\)](#) and not taken into account in the nondimensionalization which assumes the same parameters for all the different laccoliths.

Half of the large mafic sills show dimensionless radius smaller than $R = 4$, not consistent with their arrest in the gravity current regime (Figure 2.5). It might suggest that these mafic sills have intruded shallower into the crust; for instance, for $d_c = 250$ m, the characteristic length scale for the mafic sill is smaller $\Lambda = 800$ m and 95% of the population show dimensionless radius larger than 4. Nevertheless, their dimensionless thickness, which should tend to a constant of order $O(1)$ according to the model, is much larger than the expected value and increases with the radius R . For a gravity current in a two dimensional geometry, the thickness is indeed expected to increase with the length of the sill, but as $L^{1/4}$ [Michaut \(2011\)](#) which is much weaker than the found value for the coefficient a of 0.76 for large mafic sill (Figure 2.5). Therefore, the model predictions toughly reconcile with the observations for large mafic sills.

2.3.2 Origin of lunar low-slope domes

Dataset

On the Moon, 13 elongated low-slope domes, located around the lunar maria, have been recently identified as potentially intrusive domes ([Wöhler et al., 2007, 2009](#)). [Wöhler et al. \(2009\)](#) used an image-based 3D reconstruction approach which relies on a combination of photoclinometry and shape from shading techniques to determine the morphometric properties of each of these lunar domes which results in a 10% error estimate on the intrusion thickness. These data have since been updated by Mélanie Thiriet, a graduate student in our laboratory, who used the high resolution of the topography obtained from the 64 ppd (~ 450 m/pixel ([Zuber et al., 2009](#))) LOLA gridded topography data downloaded from the planetary data system geosciences

node to reevaluate the thickness and the radius of some of these potentially intrusive lunar domes.

Range of value for the parameters

Given the basalt composition of most lunar rocks from the mare brought back from the Apollo missions, the lunar magmas are more likely to be mafic in composition and we use $\rho_m = 2900 \text{ kg m}^{-3}$ for the lava density. Therefore, on the Moon, the larger lava density and the smaller gravity leads to length scale 1.5 times larger than terrestrial ones; for instance, using $E = 10 \text{ GPa}$ and $d_c = 1.5 \text{ km}$, the characteristic length scale for a lunar intrusion is $\sim 5 \text{ km}$ and 3.3 km for a terrestrial laccolith.

The source of magma in the lunar interior is poorly constrained and more likely to be deeper than on Earth; most of the mare basalt are thought to be a product of melting initiated deep in the lunar mantle, deeper than 400 km (*Shearer, 2006*). Using the same value for the initial driving pressure, $\Delta P = 50 \text{ MPa}$, unless lunar magmas are likely to be more mafic and contain less volatiles implying smaller driving pressure, and a depth of 500 km for the magma source region, the overpressure gradient is only of 100 Pa m^{-1} . Mare basalt, which have lower concentration in alkalies than terrestrial basalt, should also have a lower viscosity. We take $\eta_h = 1 \text{ Pa s}$ and for a dyke 10 m thick, the injection rate is $Q_0 = 10^4 \text{ m}^3 \text{ s}^{-1}$ and the typical height scale $H \sim 2 \text{ m}$.

Predictions versus observations

After nondimensionalization, the lunar low-slope domes show dimensionless radius smaller than 4 consistent with their arrest in the bending regime. In addition, they are almost perfectly aligned with the terrestrial laccolith from Elba Island (Figure 2.5) (*Michaut, 2011*). Indeed, the best fit for all the observations, lunar domes + Elba Island laccoliths, is $h_0 = 10R^{1.11}$ with a high correlation coefficient $r^2 = 0.94$. Given that the same intrusion depth has been arbitrarily chosen for all intrusions, the fit is surprisingly accurate. Therefore, the isoviscous elastic-plated gravity current model supports the intrusive origin of the lunar domes described by *Wöhler et al. (2009)* and their arrest in the bending regime. In addition, it is able to explain the difference between Earth laccolith and lunar intrusive domes (*Michaut, 2011*).

2.3.3 Discussion

Historical models for intermediate scale magmatic intrusion consider that the main phase of laccolith growth and spreading require a two-stage process: horizontal spreading of a sill followed by vertical inflation when the sill has grown horizontally enough so that the magma has enough leverage on the overlying layer to begin to bend them upward (*Johnson and Pollard, 1973; Koch et al., 1981*). More recent models proposed instead that these intrusions form as a series of sub-horizontally staked magma sheets (*Morgan et al., 2008; Menand, 2011*). While both model are able to account for several geological observations, they both lack a physical description of the intrusion process and are then not able to explain the solidified morphology of these magmatic intrusions in term of the flow parameters (injection rate, volume) at the time of emplacement.

Isoviscous elastic-plated gravity current

Michaut (2011) has developed a new approach to model intermediate-scale intrusions such as sills, laccoliths or bysmaliths through a dynamic elastic-plated gravity current model that consider both the bending and the own weight of the magma as driving the flow. This model shows promising results in predicting the variety of shape of intermediate scale magmatic intrusions; from the dome shape of laccolith to the disk-like morphology of large mafic sill. It allows to relate the laccolith morphology to the crustal and magma physical properties, and more importantly, to the injection rate. The prediction of the model, especially the scaling law for the thickness versus length in the bending regime, also fits the variability in the laccolith units at Elba Island, hence providing for a physical explanation for the observed relationship between laccolith length and thickness. Finally, the model shows promising results in explaining the discrepancy between terrestrial laccolith and low-slope lunar domes on the Moon and therefore can be used to assess the intrusive origin of intrusive candidates on other terrestrial planets.

However, there are questions that remain open. First, we have shown that the model hardly account for the absolute final value for both the thickness and the radius of these laccoliths and that reconciliating predictions and observations requires abnormally high magma viscosity. Second, the model was also not able to account for the increase in large mafic sill thickness with their diameter. More importantly, the model crucially lack of a stopping criterion that could predict the final morphology of a magmatic intrusion given its input parameters. Fracturation is generally considered as the limiting mechanism to the expansion of magmatic intrusions.

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Criterion for fracturation

As the flow length increases, the pressure in the intrusion eventually decreases to the critical value equal to the pressure necessary for fracturing the tip. In that case, fracturing at the tip might limit spreading and trigger the arrest of a laccolith in the bending regime. The stress intensity factor K_I for a mode I fracture and a uniformly loaded crack situated close to a boundary (i.e., $d \ll R$) can be approximated by (*Dyskin et al.*, 2000; *Bunger and Emmanuel*, 2005)

$$K_I = K_M M_0 d_c^{-3/2} \quad (2.28)$$

where $K_M = 1.932$ is a constant and M_0 is the bending moment at the crack tip given by

$$M_0 = -D \left(\frac{\partial^2 h}{\partial r^2} + \frac{1}{r} \frac{\partial h}{\partial r} \right) \Big|_{r=R(t)}. \quad (2.29)$$

Once K_I reaches the fracture toughness limit K_c , i.e. $\sim 4 - 400$ MPa m^{1/2} (*Olson*, 2003), fracturing at the tip will limit the intrusion lateral extent.

Injecting the scaling for the thickness h_0 as a function of the radius R (2.27) with dimension into the predicted flow shape in the bending regime (3.49) gives the flow shape as a function of the radius $R(t)$ of the laccolith

$$h(r, t) = 20H\Lambda^{-8/7}R(t)^{8/7} \left(1 - \frac{r^2}{R^2(t)} \right)^2. \quad (2.30)$$

Injecting this expression into (2.29) and (2.28) and inverting for the radius, one can then find that the critical dimensionless radius R_{cr} for the laccolith is

$$R_{cr} \sim \frac{20E^{7/6}H^{7/6}K_m^{7/6}}{K_c^{7/6}\Lambda^{7/3}}d^{7/4} \quad (2.31)$$

which, in terms of the parameters, reads

$$R_{cr} \sim 142E^{7/12}K_m^{7/6}Q_0^{7/24}\eta_h^{7/24}g^{7/24}\rho_m^{7/24}K_c^{-7/6} \quad (2.32)$$

and therefore depends mainly in the fracture toughness of the encasing rocks. One can calculate that for typical crustal and magma parameters taken in Section 2.3 for terrestrial laccoliths and the largest reported value for the parameter $K_c = 4000$ MPa m^{1/2}, the critical dimensionless radius is equal to ~ 10 and therefore, still much larger than the transition radius between the bending and gravity regime $R = 4$. Therefore, while fracturation might explain the arrest of magmatic intrusion for large mafic sill, it does not provide a sufficient mechanism for the arrest of laccolith.

2.4 Toward a more realistic model for shallow magmatic intrusions

We have shown that while it shows promising results in reproducing the morphology of solidified magmatic intrusions on Earth and the Moon, the isoviscous elastic-plated gravity current suffers some limitations as well. In this manuscript, we propose to explore two important mechanisms that have been neglected until now and will certainly influence the emplacement of shallow magmatic intrusions in the crust of terrestrial planets: the effect a temperature-dependent rheology for the magma and the effect of an overburden characterized by a non-constant thickness.

The former has already shown important implications for the cooling of lava domes (*Bercovici, 1994; Bercovici and Lin, 1996; Balmforth and Craster, 2004; Garel et al., 2014*). Indeed, the viscosity of magma can vary by several orders of magnitudes during the cooling (*Shaw, 1972; Lejeune and Richet, 1995*) and therefore, as the fluid cools, its composition and crystal content change which, in turn, modifies the viscosity and the dynamics of the flow itself. The first part of the manuscript deals with this matter and try to better understand the dynamics of a cooling elastic-plated gravity current. In particular, in chapter 3, we derive a model of a cooling elastic-plated gravity current with a viscosity inversely dependent on temperature and isothermal boundary conditions. This model is next further refined and compared to the observation in chapter 4.

The second part of the manuscript addresses the second point and in particular, the problem of crater-centered intrusions with application to the endogenous deformation observed at lunar floor-fractured craters. Indeed, if large volume of magma have effectively intruded at shallow depth below these craters as proposed by *Schultz (1976)*, the topography of the initial impact crater has to constrain the underlying magma spreading dynamics. Chapter 5 presents a model for crater-centered intrusions and its application to the deformations observed at floor-fractured craters. Finally, chapter 6 takes the study of floor-fractured craters one step further by looking at the gravitational signature of lunar floor-fractured craters in the light of the model predictions.

This approach, evolving each time from a theoretical model to the observations, might help shed light both on the limitations of the model of *Michaut (2011)* exposed in this chapter and the origin of lunar-floor fractured craters.

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Part II

Evolution thermique des intrusions magmatiques à faible profondeur

CHAPTER 3

Elastic-plated gravity current with temperature-dependent viscosity

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Temperature-dependent elastic-plated gravity current has numerous applications in the nature science, from the intrusion of magma is the shallow layer of the crust to the flowing of melt-water below ice sheet. We develop the general equations for the elastic-plated gravity current with temperature-dependent viscosity for constant influx conditions. Crystallization is also taken into account as a source/sink of heat when the fluid crystallizes/melts during emplacement. We show that the coupling between the thermal structure and the flow itself results in important deviations from the isoviscous case. In particular, both regimes, taken individually, split in three phases. In the first phase, the thermal anomaly has the size of the current itself, the effective viscosity is minimal and the current spreads as in the isoviscous case. The second phase is triggered by the detachment of the thermal anomaly and followed by an important increase in the effective viscosity of the flow. The current slows down and thickens. Finally, when the cold front region becomes about 10% of the flow itself, the effective viscosity stabilizes to its maximum value and the current return in an isoviscous dynamics, but with cold viscosity. Further analyses show that the effective viscosity is the average viscosity of a small region a the front of the current in the bending regime and the average viscosity of the current in the gravity regime. Therefore, the evolution of an elastic-plated gravity current depends on the relative phase changes within the two regime and the transition between the two regime itself. Application to terrestrial laccolith, which is a major process in the crust formation, show that they should preferentially solidify in the third phase of the bending regime.

3.1 Introduction

Elastic-plated gravity currents involve the spreading of viscous material beneath an elastic sheet. The applications range from the emplacement of lava in the shallow crust (*Michaut*, 2011; *Bunger and Cruden*, 2011) and melt-water drainage below ice sheet (*Das et al.*, 2008; *Tsai and Rice*, 2010) in geological setting to the manufacture of flexible electronics and microelectromechanical systems (MEMS) in engineering (*Hosoi and Mahadevan*, 2004).

When the thickness of the flow is small compared to its extent, lubrication approximation applies and the study of elastic-plated gravity currents resumes to the study of a sixth order, non-linear partial differential equation (*Michaut*, 2011; *Lister et al.*, 2013; *Hewitt et al.*, 2014). However, the assumption that the thickness of the fluid tends to zero at the contact line leads to divergent viscous stresses, and hence, the need of a regularization condition at the front (*Flitton and King*, 2004; *Lister et al.*, 2013; *Hewitt et al.*, 2014). One common approach is to add a thin pre-wetted film of fluid, thus avoiding the requirement for any boundary conditions at a genuine contact line (*Lister et al.*, 2013; *Hewitt et al.*, 2014).

The dynamics of the spreading has been described in an axisymmetric geometry for a Newtonian fluid with constant viscosity (*Michaut*, 2011; *Lister et al.*, 2013; *Thorey and Michaut*, 2014). In particular, they show the presence of two distinct regimes of evolution. First, gravity is negligible and the peeling of the front is driven by bending; the interior is bell-shaped, the radius evolves as $t^{8/22}$ and the thickness evolves as $t^{7/22}$. When the radius becomes larger 4Λ , where Λ is the flexural wavelength of the upper layer, the weight of the current becomes dominant over the bending terms and the evolution enters a gravity current regime (*Huppert*, 1982a). In this second regime, the thickness profile shows a flat top with bent edges, the radius evolves as $t^{1/2}$ and the thickness tends to a constant. Different analogue experiments of isoviscous flows confirm these theoretical results (*Dixon and Simpson*, 1987; *Lister et al.*, 2013).

However, in many real geological settings, the isothermal/isoviscous assumption are not valid. Indeed, many geological elastic-plated gravity currents are comprised of fluid whose viscosity can vary by several orders of magnitude depending on its temperature. This is the case for magmas produced by partial melting of the upper mantle and intruding the shallow layers of the crust (*Shaw*, 1972; *Lejeune and Richey*, 1995). Therefore, as the fluid is cooling, its composition and crystal content changes which, in turn, modifies the viscosity and the dynamics of the flow. Several studies have shown that this coupling between the cooling and the flow itself in a gravity current results in important deviations from the isoviscous case (*Bercovici*, 1994; *Balmforth*

and Craster, 2004; Garel et al., 2014).

In this paper, we examine how the spreading of an elastic-plated gravity current is affected by the cooling itself. In particular, we consider the problem of an elastic-plated gravity current whose viscosity depends on temperature according to a prescribed rheology $\eta(T)$. This gives rise to a set of two coupled non-linear equations that we solve numerically. We study the flow thermal structure and its effect on the dynamics through the rheology in each regime separately. In both regimes, we identify different “thermal” phases of propagation that we characterize by different scaling laws. We then discuss our result implications concerning the emplacement of terrestrial laccoliths.

3.2 Theory

3.2.1 Formulation

We model the axisymmetric flow of fluid below an elastic layer of constant thickness d_c and above a semi infinite rigid layer (*Michaut, 2011*) (Figure 3.1). The assumption that the thickness of the fluid $h(r, t)$ tends to zero at the contact line leads to divergent viscous stresses and to the theoretical immobility of the current (*Flitton and King, 2004*). To avoid problem at the contact line, we consider a thin pre-wetting film of thickness h_f (*Lister et al., 2013*) (Figure 3.1).

The fluid is injected continuously at the base and center of the current at a constant rate Q_0 through a conduit of diameter a . The hot fluid is intruded at temperature T_i and cools through the top and bottom by conduction in the surrounding medium, whose temperature is considered constant and equal to T_0 . In using a fixed temperature at the flow boundary, we essentially assume the fluid is bounded by a medium with infinite thermal conductivity.

As it cools, the viscosity of the fluid increases following a prescribed rheology $\eta(T)$ given by

$$\eta(T) = \frac{\eta_h \eta_c (T_i - T_0)}{\eta_h (T_i - T_0) + (\eta_c - \eta_h)(T - T_0)} \quad (3.1)$$

where η_h and η_c are the viscosities of the hottest and coldest fluid at the temperature T_i and T_0 respectively (*Bercovici, 1994*). Although this rheology is largely simplified, the inverse dependence of viscosity on temperature captures the essential behavior of a viscous fluid, i.e. the viscosity variations are the largest where the temperature is the coldest (*Shaw, 1972; Marsh, 1981; Lejeune and Richet, 1995; Giordano et al., 2008*).

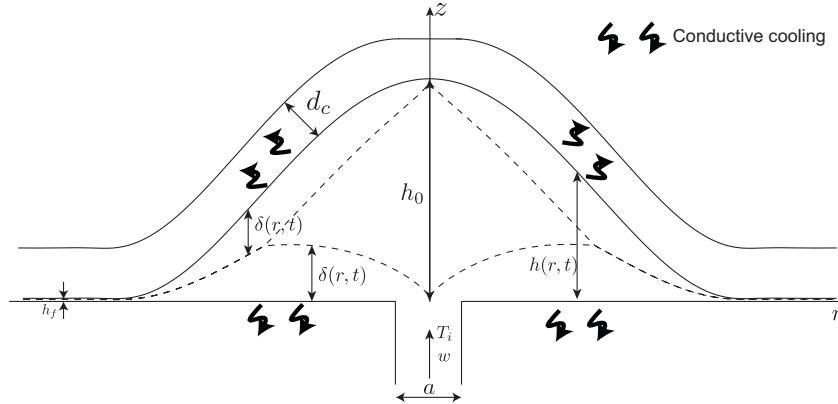


Figure 3.1: Model geometry and parameters. The vertical scale is exaggerated.

3.2.2 Pressure

The intrusion develops over a length scale Λ that is much larger than its thickness H ($\Lambda \gg H$). In the laminar regime and in axisymmetrical coordinates (r, z) , the Navier-Stokes equations under the lubrication assumption are

$$-\frac{\partial P}{\partial r} + \frac{\partial}{\partial z} \left(\eta(T) \frac{\partial u}{\partial z} \right) = 0 \quad (3.2)$$

$$-\frac{\partial P}{\partial z} - \rho_m g = 0 \quad (3.3)$$

where $u(r, z, t)$ is the radial velocity, ρ_m the fluid density, g the standard acceleration due to gravity and $P(r, z, t)$ the pressure within the fluid. Integration of (3.3) gives the total pressure $P(r, z, t)$ within the flow. When the vertical deflection $h(r, t)$ of the upper elastic layer is small compared to its thickness d_c , i.e $h \ll d_c$, we can neglect stretching of the upper layer and only consider bending stresses. Therefore, the total pressure $P(r, z, t)$ at a level z in the current is the sum of three contributions: the weight of the magma and of the upper layer and the bending pressure

$$P = \rho_m g(h - z) + \rho_r g d_c + D \nabla_r^4 h \quad (3.4)$$

where $h(r, t)$ is the flow thickness, ρ_r the density of the surrounding rocks and D is the flexural rigidity of the thin elastic layer, that depends on Young's modulus E , Poisson's ratio ν^* and on the elastic layer thickness d_c as $D = E d_c^3 / (12(1 - \nu^*))$.

3.2.3 Injection rate

Assuming a Poiseuille flow within the cylindrical feeding conduit, the vertical injection velocity $w_i(r, t)$ and injection rate Q_0 are given by

$$w_i(r, t) = \begin{cases} \frac{\Delta P}{4\eta_h Z_c} \left(\frac{a^2}{4} - r^2 \right) & r \leq \frac{a}{2} \\ 0 & r > \frac{a}{2} \end{cases} \quad (3.5)$$

$$Q_0 = \frac{\pi \Delta P a^4}{128 \eta_h Z_c} \quad (3.6)$$

where ΔP is the initial overpressure within the melt at $z = Z_c$.

3.2.4 Heat transport equation

3.2.4.1 Local energy conservation

In the laminar regime and in axisymmetrical coordinates (r, z) , the local energy conservation equation within the lubrication assumption is written as

$$\frac{D}{Dt} (\rho_m C_{p,m} T + \rho_m L(1 - \phi)) = k_m \frac{\partial^2 T}{\partial z^2} \quad (3.7)$$

where $T(r, z, t)$ is the fluid temperature and ρ_m , k_m and $C_{p,m}$ are the density, thermal conductivity and specific heat of the fluid. Here, we also account for energy release by crystallization of the fluid, which is a non negligible source of heat for magmas; $\phi(r, z, t)$ is the crystal fraction in the melt and L the latent heat. In this model, the crystals are considered only as a source/sink of energy as they melt/form during flow emplacement. In particular, the physical properties of the fluid are not modified by the presence of crystals.

Following a common approximation, we assume that the crystal fraction is a linear function of temperature over the melting interval

$$\phi = \frac{T_L - T}{T_L - T_s} \quad (3.8)$$

where T_s and T_L are the solidus and liquidus temperatures of the magma (*Hort*, 1997; *Michaut and Jaupart*, 2006). In addition, we assume that the fluid is injected as its liquidus temperature, i.e. $T_L = T_i$ and, for simplicity, that the solidus temperature is equal to the surrounding rock temperature $T_s = T_0$. With these approximations, the local energy equation (3.7) resumes to

$$\frac{\partial T}{\partial t} + u \frac{\partial T}{\partial r} + w \frac{\partial T}{\partial z} = \frac{St}{St + 1} \kappa_m \frac{\partial^2 T}{\partial z^2} \quad (3.9)$$

where $u(r, z, t)$ and $w(r, z, t)$ are the radial and vertical fluid velocities, $St = (C_{p,m}(T_i - T_0)) / L$ is the Stephan number and κ_m is the fluid thermal diffusivity $\kappa_m = k_m / (\rho_m C_{p,m})$. We use an integral balance method to solve the heat transport equation (3.9). This theory is based on the integral-balance method of heat-transfer theory of [Goodman \(1958\)](#), in which the vertical structure of the temperature field is represented by a known function of depth that approximates the expected solution.

3.2.4.2 Integral balance solution for the temperature $T(r, z, t)$

Following [Balmforth and Craster \(2004\)](#), we model the cooling of the flow through the growth of two thermal boundary layers: one growing downward from the top and a second growing upward from the base. As we consider homogeneous thermal properties for the surrounding rocks, we assume that the two thermal boundary layers grow symmetrically and have the same thickness $\delta(r, t)$. We use the following approximation for the vertical temperature profile $T(r, z, t)$

$$T = \begin{cases} T_b - (T_b - T_0)(1 - \frac{z}{\delta})^2 & 0 \leq z \leq \delta \\ T_b & \delta \leq z \leq h - \delta \\ T_b - (T_b - T_0)(1 - \frac{h-z}{\delta})^2 & h - \delta \leq z \leq h \end{cases} \quad (3.10)$$

where $T_b(r, t)$ is the temperature at the center of the flow. The integral balance solution in (3.10) assumes a symmetry around $z = h/2$ and a decrease of the temperature in the two thermal boundary layers down to the surrounding rock temperature T_0 ([Balmforth and Craster, 2004](#)). In addition, it assumes a uniform temperature T_b in between the thermal boundary layers. As the fluid is injected at temperature T_i , we have $T_b(r, t) = T_i$ as long as $\delta < h/2$. However, if the two thermal boundary layers connect, then $\delta = h/2$ and T_b becomes such that $T_b \leq T_i$. This profile assures the continuity of the temperature and heat flux within the flow.

3.2.4.3 Integral balance equation

We begin by integrating the local energy conservation equation (3.9) separately over the two thermal boundary layers. The integration over the bottom thermal layer, i.e. from the base, $z = 0$ to a level $z = \delta$ gives

$$\begin{aligned} & \frac{\partial}{\partial t} (\delta(\bar{T} - T_b)) + \frac{1}{r} \frac{\partial}{\partial r} (r\delta(\bar{u}\bar{T} - \bar{u}T_b)) + \delta \left(\frac{\partial T_b}{\partial t} + \bar{u} \frac{\partial T_b}{\partial r} \right) \\ &= -\frac{\kappa_m}{1 + St} \frac{\partial T}{\partial z} \Big|_{z=0} + w_i(T_i - T_b) \end{aligned} \quad (3.11)$$

where the bars indicate the vertical average over the bottom thermal boundary layer

$$\bar{f} = \frac{1}{\delta} \int_0^\delta f dz,$$

$T_b(r, t)$ is the temperature at $z = \delta$, $w_i(r)$ is the vertical injection velocity and we have used the nullity of the thermal gradient at $z = \delta$ and the local mass conservation

$$\frac{1}{r} \frac{\partial r u}{\partial r} + \frac{\partial w}{\partial z} = 0. \quad (3.12)$$

The integration over the top thermal layer, i.e., from the level, $z = h - \delta$ to the top $z = h$ gives

$$\begin{aligned} & \frac{\partial}{\partial t} (\delta(\bar{T} - T_b)) + \frac{1}{r} \frac{\partial}{\partial r} (r \delta(\bar{u}T - \bar{u}T_b)) + \delta \left(\frac{\partial T_b}{\partial t} + \bar{u} \frac{\partial T_b}{\partial r} \right) \\ &= \frac{\kappa_m}{1 + St^{-1}} \frac{\partial T}{\partial z} \Big|_{z=h}. \end{aligned} \quad (3.13)$$

where, in addition to the local mass conservation (3.12) and the fact that the thermal gradient at $z = h - \delta$ is equal to zero, we have used the kinematic boundary condition in $z = h(r, t)$

$$\frac{\partial h}{\partial t} + u \frac{\partial h}{\partial r} = w \quad (3.14)$$

Therefore, the heat balance equation, i.e. the heat equation (3.9) integrated over the flow thickness, is obtained by adding (3.11) and (3.13). Introducing (3.10) to derive the conductive fluxes, we finally obtain

$$\begin{aligned} & \frac{\partial}{\partial t} (\delta(\bar{T} - T_b)) + \frac{1}{r} \frac{\partial}{\partial r} (r \delta(\bar{u}T - \bar{u}T_b)) + \delta \left(\frac{\partial T_b}{\partial t} + \bar{u} \frac{\partial T_b}{\partial r} \right) \\ &= -\frac{2\kappa_m}{(1 + St^{-1})} \frac{(T_b - T_0)}{\delta} + \frac{w_i}{2} (T_i - T_b) \end{aligned} \quad (3.15)$$

3.2.5 Equation of motion

A global statement of mass conservation gives

$$\frac{\partial h}{\partial t} + \frac{1}{r} \frac{\partial}{\partial r} \left(r \int_0^h u dz \right) = w_i. \quad (3.16)$$

To obtain an equation for the flow thickness, we first note that the chosen vertical structure of the temperature field (3.10) is symmetric around $h/2$,

and thus, the viscosity and velocity u possess the same symmetry. Taking advantage of this symmetry, we integrate once (3.2) using $\frac{\partial u}{\partial z}|_{z=h/2} = 0$ to get

$$\frac{\partial u}{\partial z} = \frac{1}{\eta} \frac{\partial P}{\partial r} \left(z - \frac{h}{2} \right). \quad (3.17)$$

Using no-slip boundary conditions at the top and the bottom of the flow, i.e. $u(r, z=0, t) = u(r, z=h, t) = 0$ equation (3.16) can be rewritten as

$$\frac{\partial h}{\partial t} = \frac{1}{r} \frac{\partial}{\partial r} \left(r \int_0^h \frac{\partial u}{\partial z} z dz \right) + w_i \quad (3.18)$$

and injecting (3.17) into (3.18) finally gives the equation for the flow thickness evolution in axisymmetric coordinates

$$\frac{\partial h}{\partial t} = \frac{1}{r} \frac{\partial}{\partial r} \left(r \left(\rho_m g \frac{\partial h}{\partial r} + D \frac{\partial}{\partial r} (\nabla_r^4 h) \right) \left(\int_0^h \frac{1}{\eta(y)} \left(y - \frac{h}{2} \right) y dy \right) \right) + w_i \quad (3.19)$$

In addition, integration of (3.17) using the no-slip boundary condition at the base of the flow gives

$$u(r, z, t) = \frac{\partial P}{\partial r} \int_0^z \frac{1}{\eta(y)} \left(y - \frac{h}{2} \right) dy. \quad (3.20)$$

3.2.6 Dimensionless equations

We use the characteristic temperature interval $\Delta T = T_i - T_0$ to nondimensionalize temperatures. The dimensionless integral balance approximation (3.10) becomes

$$\theta(z) = \begin{cases} \Theta_b \left(1 - (1 - \frac{z}{\delta})^2 \right) & 0 \leq z \leq \delta \\ \Theta_b & \delta \leq z \leq h - \delta \\ \Theta_b \left(1 - (1 - \frac{h-z}{\delta})^2 \right) & h - \delta \leq z \leq h \end{cases} \quad (3.21)$$

where $\theta(r, z, t)$ is the dimensionless temperature and $\Theta_b = \frac{T_b - T_0}{T_i - T_0}$. Finally, equations (3.15) and (3.19) are nondimensionalized using a horizontal scale Λ , a vertical scale H and a time scale τ given by

$$\Lambda = \left(\frac{D}{\rho_m g} \right)^{1/4} \quad (3.22)$$

$$H = \left(\frac{12 \eta_h Q_0}{\rho_m g \pi} \right)^{1/4} \quad (3.23)$$

$$\tau = \frac{\pi \Lambda^2 H}{Q_0} \quad (3.24)$$

Chapter 3. Elastic-plated gravity current with temperature-dependent viscosity

where Λ represents the flexural wavelength of the upper elastic layer (*Michaut*, 2011), H the characteristic thickness of an isoviscous constant flux gravity current with viscosity η_h (*Huppert*, 1982b) and τ the characteristic time to fill up a cylindrical flow of radius Λ and thickness H at a constant rate Q_0 . In addition, we can define a horizontal velocity scale $U = \Lambda/\tau = (\rho_m g H^3) / (12\eta_h \Lambda)$.

The dimensionless equations are

$$\frac{\partial h}{\partial t} = \frac{12}{r} \frac{\partial}{\partial r} \left(r \left(\frac{\partial h}{\partial r} + \frac{\partial}{\partial r} (\nabla_r^4 h) \right) I_1(h) \right) + w_i \quad (3.25)$$

$$\begin{aligned} \frac{\partial}{\partial t} (\delta(\bar{\theta} - \Theta_b)) &= -\frac{1}{r} \frac{\partial}{\partial r} (r\delta(\bar{u}\theta - \bar{u}\Theta_b)) - \delta \left(\frac{\partial \Theta_b}{\partial t} + \bar{u} \frac{\partial \Theta_b}{\partial r} \right) \\ &\quad - 2Pe^{-1} St_m \frac{\Theta_b}{\delta} + \frac{w_i}{2}(1 - \Theta_b) \end{aligned} \quad (3.26)$$

$$w_i = \frac{32}{\gamma^2} \left(\frac{1}{4} - \frac{r^2}{\gamma^2} \right) \text{ if } r < \gamma/2, \quad w_i = 0 \text{ if } r \geq \gamma/2 \quad (3.27)$$

$$u(r, z, t) = 12 \left(\frac{\partial h}{\partial r} + \frac{\partial}{\partial r} (\nabla_r^4 h) \right) I_0(z) \quad (3.28)$$

with

$$I_0(z) = \int_0^z (\nu + (1 - \nu)\theta(y)) \left(y - \frac{h}{2} \right) dy \quad (3.29)$$

$$I_1(z) = \int_0^z (\nu + (1 - \nu)\theta(y)) \left(y - \frac{h}{2} \right) y dy \quad (3.30)$$

and where γ , Pe , St_m and ν are the four dimensionless numbers that control the dynamics of the flow

$$\gamma = \frac{a}{\Lambda} \quad (3.31)$$

$$Pe = \frac{H^2}{\kappa_m \tau} \quad (3.32)$$

$$St_m = \frac{C_{p,m} (T_i - T_0)}{C_{p,m} (T_i - T_0) + L} \quad (3.33)$$

$$\nu = \frac{\eta_h}{\eta_c}. \quad (3.34)$$

γ is the dimensionless radius of the conduit, it does not significantly influence the flow and is set to 0.02 in this study (*Michaut and Bercovici*, 2009; *Michaut*, 2011); Pe is the Peclet number which compares the vertical diffusion of heat to the horizontal advection in the interior; St_m is a modified Stephan number which represents the ratio of sensible heat between solidus and liquidus to the total energy of the fluid at liquidus temperature and ν is the maximum viscosity contrast, i.e. the ratio between the hottest and coldest viscosity.

3.2.7 Further simplifications

3.2.7.1 Heat equation

The heat balance equations (3.26) can reduce to

$$\frac{\partial}{\partial t} (\delta(\bar{\theta} - 1)) + \frac{1}{r} \frac{\partial}{\partial r} (r\delta(\bar{u}\theta - \bar{u})) = -2Pe^{-1}St_m \frac{\Theta_b}{\delta} \quad (3.35)$$

Indeed, if the thermal boundary layers exist, $\Theta_b = 1$, δ is the variable and (3.26) directly reduces to (3.35). In contrast, if the thermal boundary layers merge, $\delta = h/2$ and the variable is Θ_b . In this case, the heat balance equation (3.26) reduces to

$$\frac{\partial h\bar{\theta}}{\partial t} + \frac{1}{r} \frac{\partial}{\partial r} (rh\bar{u}\theta) - \Theta_b \left(\frac{\partial h}{\partial t} + \frac{1}{r} \frac{\partial}{\partial r} (rh\bar{u}) \right) = -8St_mPe^{-1} \frac{\Theta_b}{h} + w_i \quad (3.36)$$

which, by using (3.16), rewrites

$$\frac{\partial h\bar{\theta}}{\partial t} + \frac{1}{r} \frac{\partial}{\partial r} (rh\bar{u}\theta) = w_i - 8St_mPe^{-1} \frac{\Theta_b}{h}. \quad (3.37)$$

Equation (3.37) also corresponds to (3.35) when $\delta = h/2$.

Following *Balmforth and Craster* (2004), we rewrite (3.35) using a new variable $\xi = \delta(1 - \bar{\theta})$

$$\frac{\partial \xi}{\partial t} + \frac{1}{r} \frac{\partial}{\partial r} (r\bar{u}\xi) - \frac{1}{r} \frac{\partial}{\partial r} (r\delta(\bar{u}\theta - \bar{u}\bar{\theta})) = 2Pe^{-1}St_m \frac{\Theta_b}{\delta}. \quad (3.38)$$

where Θ_b and δ can be calculated directly from the expression of ξ such that

$$\Theta_b(r) = \begin{cases} 1 & \text{if } \xi \leq \xi_t \\ \frac{3}{2} - \frac{3\xi}{h} & \text{if } \xi > \xi_t \end{cases} \quad \delta(r) = \begin{cases} 3\xi & \text{if } \xi \leq \xi_t \\ h(r,t)/2 & \text{if } \xi > \xi_t \end{cases}$$

with $\xi_t = h/6$.

The second term on the left hand side of (3.38) contains advection by the vertically integrated radial velocity while the third term contains a correction accounting for the vertical structure of the temperature field. The term on the right is the loss of heat by conduction in the surrounding medium.

3.2.7.2 Average quantities

The average velocity over a thermal boundary layer \bar{u} reads

$$\bar{u} = \frac{1}{\delta} \int_0^\delta u dz = u(r, \delta, t) - \frac{1}{\delta} \int_0^\delta \frac{\partial u}{\partial z} z dz \quad (3.39)$$

$$= \frac{12}{\delta} \frac{\partial P}{\partial r} (\delta I_0(\delta) - I_1(\delta)) \quad (3.40)$$

where $P(r, z, t) = h + \nabla_r^4 h$ is the dimensionless pressure and we have used (3.17) in (3.39). The average rate of heat advected $\bar{u}\theta$ over a thermal boundary layer reads

$$\begin{aligned} \bar{u}\theta &= \frac{1}{\delta} \int_0^\delta u\theta dz = \frac{1}{\delta} \left([uG(z)]_0^\delta - \int_0^\delta G(z) \frac{\partial u}{\partial z} dz \right) \\ &= \frac{12}{\delta} \frac{\partial P}{\partial r} (G(\delta)I_0(\delta) - I_2(\delta)) \end{aligned} \quad (3.41)$$

where

$$G(z) = \Theta_b \left(z + \frac{\delta}{3} \left(1 - \frac{z}{\delta} \right)^3 \right) \quad (3.42)$$

denotes a primitive of θ when $z < \delta$ and

$$I_2(z) = \int_0^y (\nu + (1-\nu)\theta(y)) G(y) \left(y - \frac{h}{2} \right) dy. \quad (3.43)$$

Therefore, we have

$$\bar{u}\theta - \bar{u}\bar{\theta} = \frac{12}{\delta} \frac{\partial P}{\partial r} (I_0(\delta) (G(\delta) - \delta\bar{\theta}) + \bar{\theta}I_1(\delta) - I_2(\delta)) \quad (3.44)$$

where the average temperature over a thermal boundary layer is $\bar{\theta} = 2\Theta_b/3$

3.2.8 Summary of the equations

The coupled equations governing the cooling of an elastic-plated gravity current are summarized in term of the integrals (3.29), (3.30) and (3.43) as follow

$$\frac{\partial h}{\partial t} - \frac{12}{r} \frac{\partial}{\partial r} \left(r I_1(h) \frac{\partial P}{\partial r} \right) = \mathcal{H} \left(\frac{\gamma}{2} - r \right) \frac{32}{\gamma^2} \left(\frac{1}{4} - \frac{r^2}{\gamma^2} \right) \quad (3.45)$$

$$\frac{\partial \xi}{\partial t} + \frac{1}{r} \frac{\partial}{\partial r} (r (\bar{u}\xi - \Sigma)) = 2Pe^{-1} St_m \frac{\Theta_b}{\delta} \quad (3.46)$$

with

$$\Theta_b(r) = \begin{cases} 1 & \text{if } \xi \leq \xi_t \\ \frac{3}{2} - \frac{3\xi}{h} & \text{if } \xi > \xi_t \end{cases} \quad \delta(r) = \begin{cases} 3\xi & \text{if } \xi \leq \xi_t \\ h(r, t)/2 & \text{if } \xi > \xi_t \end{cases}$$

$$\bar{u} = \frac{12}{\delta} \frac{\partial P}{\partial r} (\delta I_0(\delta) - I_1(\delta)) \quad (3.47)$$

$$\Sigma = \frac{\partial P}{\partial r} (8I_1(\delta)\Theta_b - 12I_2(\delta)) \quad (3.48)$$

$P = h + \nabla_r^4 h$ is the dimensionless pressure and \mathcal{H} the Heaviside function. The expression of $I_0(\delta)$, $I_1(h)$, $I_1(\delta)$ and $I_2(\delta)$ as well as the numerical scheme are given in appendix 3.8.

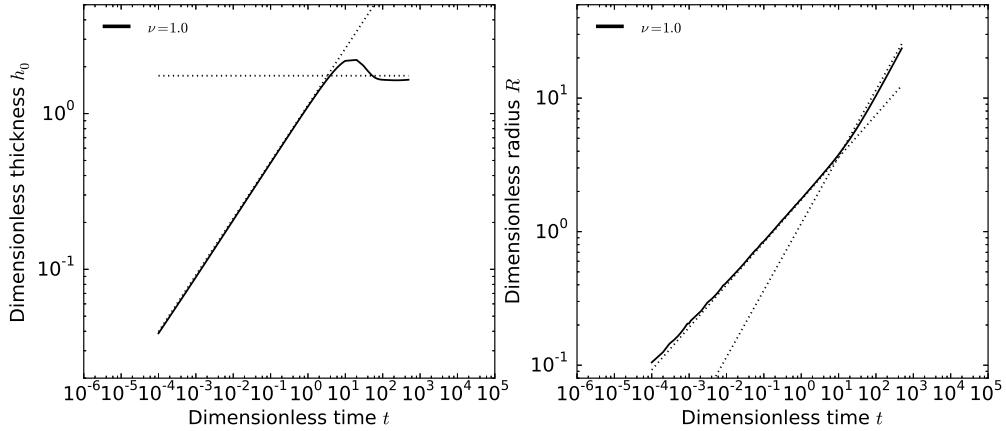


Figure 3.2: Left: Dimensionless thickness at the center h_0 versus dimensionless time t . Dotted-lines: scaling laws in the bending regime $h_0 = 0.7h_f^{-1/11}t^{8/22}$ and the gravity regime h_0 tends to a constant. Right: Dimensionless radius R versus dimensionless time t . Dotted-lines: scaling laws in the bending regime $R = 2.2h_f^{1/22}t^{7/22}$ and in the gravity current regime $R \propto t^{1/2}$.

3.2.9 Preliminary results for an isothermal flow

For a constant injection rate, a small pre-wetting film thickness, i.e. $h_f \ll 1$ and a viscosity contrast ν set to 1, numerical resolution of (3.45) shows two asymptotic spreading regimes ([Michaut, 2011](#); [Lister et al., 2013](#)).

At early times, when $R \ll \Lambda$, gravity is negligible and the spreading dynamics is governed by the bending of the upper layer. The spreading is very slow and the interior has uniform pressure $P = \nabla_r^4 h$. The flow is bell-shaped and its thickness is given by

$$h(r, t) = h_0(t) \left(1 - \frac{r^2}{R^2(t)}\right)^2 \quad (3.49)$$

with $h_0(t)$ the thickness of the current at the center ([Michaut, 2011](#); [Lister et al., 2013](#)). In this regime, [Lister et al.](#) (2013) have shown that the spreading is controlled by the propagation of a peeling by bending wave at the flow front with dimensionless velocity c

$$c = \frac{\partial R}{\partial t} = h_f^{1/2} \left(\frac{\kappa}{1.35}\right)^{5/2} \quad (3.50)$$

where $\kappa = \partial^2 h / \partial r^2$ is the dimensionless curvature of the interior solution. Using the propagation law (3.50) and the form of the interior solution (3.49),

they find that the flow radius and height are given by the following solutions

$$h_0(t) = 0.7 h_f^{-1/11} t^{8/22} \quad (3.51)$$

$$R(t) = 2.2 h_f^{1/22} t^{7/22}. \quad (3.52)$$

where the numerical pre-factor obtained in our simulations match those of *Lister et al.* (2013) (Figure 3.2).

In contrast, when the radius R becomes larger than 4Λ ($R \gg \Lambda$), the weight of the current becomes dominant over the bending terms. The pressure is given by the hydrostatic pressure $P = h$ and the current enters a classical gravity current regime where bending terms only affect the solution near the edge of the current (Huppert, 1982a; Michaut, 2011; Lister et al., 2013). In this second regime, the radius evolves as $t^{1/2}$ and the thickness tends to a constant (Figure 3.2).

In the following, we study the effect of the cooling on the flow dynamics in both regimes separately. We first describe the thermal structure for an isoviscous flow, i.e. $\nu = 1$ and then study the effect of the temperature-dependent viscosity on the flow dynamics without crystallization, i.e $St_m = 1$. Finally, we look at the effect of crystallization by setting $St_m < 1$. For simplicity, we present the results for a given film thickness ($h_f = 5 \cdot 10^{-3}$). Results for different film thicknesses are shown in Appendix 3.8.

3.3 Numerical approach

3.3.1 General procedure

The coupled nonlinear partial differential equations (3.45) and (3.46) are solved on a grid of size M defined by the relation $r_i = (i - 0.5)\Delta r$ for $i = 1, \dots, M$. The grid is shifted at the center to avoid problem arising from the axisymmetrical geometry. We index the grid point by the indice i and denote the solution on this grid h_i and ξ_i and the secondary variables $\Theta_{b,i}$, $\Theta_{s,i}$ and δ_i . Both equations can be expressed on the convenient form

$$\frac{\partial u}{\partial t} - f = 0 \quad (3.53)$$

where u is the function we want to integrate and f a non-linear function that depends on u . We solve these equations by first discretizing all the spatial derivatives using Finite Difference. The accuracy of the scheme is determined by the higher order derivatives since their numerical approximation requires the largest number of sample points. We then get two systems of M ordinary

differential equations with the form

$$\frac{\partial u_i}{\partial t} - f_i = 0 \quad i = 1, \dots, M \quad (3.54)$$

The time derivatives are first order and, since explicit schemes tend to be very sensitive and unstable, we use a fully implicit backward Euler scheme to get

$$\frac{u_i^{n+1} - u_i^n}{\Delta t} - f_i(u_i^{n+1}) = 0 \quad i = 1, \dots, M \quad (3.55)$$

Since $f_i(u_i^{n+1})$ is not a linear function, the system above cannot be re-arranged to solve u_i^{n+1} in term of u_i^n and an iterative method has to be employed instead. Fixed point iteration method have shown poor results in converging toward the solution and we finally apply second order Newton's method to obtain the solution at each time step. In particular, we first linearize u^{n+1} around a guess of the solution by assuming $u^{n+1} = u^* + \delta u^n$, where u^* is a guess and δu^n is the error and we drop the i for clarity. Then, we expressed the non-linear part using a Taylor's expansion

$$f^{n+1} = f(u^{n+1}) = f(u^* + \delta u^n) = f(u^*) + J_f^h(u^*)\delta u^n$$

where $J_f^h(u^*)$ is the jacobian matrix for the function f evaluated in h^* . Injecting the expansion into (3.55) finally gives a system of M linear equations for the correction term δ_h^n which can be expressed as

$$(I - \Delta t J_f^h(u^*))\delta u^n = u^n - u^* + \Delta t f(u^*) \quad (3.56)$$

where I is the identity matrix. Therefore, each iteration solves for δu^n and we use $u_n + \delta u^n$ as a new guess u^* in each iteration. This is repeated until δu^n becomes sufficiently small. Finally, since the equations are coupled, we use a fixed-point iteration method to converge toward the solution (h, ξ) at each time step. Therefore, the algorithm is the following at each time step

- Start with a guess for the values of all variables.
- Solve the thickness equation (3.45) for h^{n+1} using Newton-Rhaphson method.
- Solve the heat equation (3.46) for ξ^{n+1} using h^{n+1} as a new guess for h^* and Newton-Rhaphson method.
- Repeat step one until further iterations cease to produce any significant changes in the values of both h^{n+1} and ξ^{n+1} .

The computational scheme is summarized in the following.

3.3.2 Thickness equation

The thickness equation (3.45) is written as

$$\frac{\partial h}{\partial t} - f(h, \xi) = 0 \quad (3.57)$$

with

$$f = \frac{1}{r} \frac{\partial}{\partial r} \left(r\phi \left(\frac{\partial}{\partial r} (h + P) \right) \right) + w_i \quad (3.58)$$

$$\phi = 12I_1(h) \quad (3.59)$$

and where P is the dimensionless bending pressure $P = \nabla^4 h$.

Spatial discretization of f

The spatial discretization is obtained using a central difference scheme over a sub-grid shifted by $0.5\Delta r$ from the main grid. Therefore, we have

$$\begin{aligned} f_i &= \frac{1}{r_i \Delta r} \left(r_{i+1/2} \phi_{i+1/2} \left(\frac{\partial h}{\partial r} + \frac{\partial P}{\partial r} \right) \Big|_{i+1/2} - r_{i-1/2} \phi_{i-1/2} \left(\frac{\partial h}{\partial r} + \frac{\partial P}{\partial r} \right) \Big|_{i-1/2} \right) \\ &= A_i \phi_{i+1/2} (h_{i+1} - h_i) - B_i \phi_{i-1/2} (h_i - h_{i-1}) \\ &\quad + A_i \phi_{i+1/2} (P_{i+1} - P_i) - B_i \phi_{i-1/2} (P_i - P_{i-1}) \\ &\quad + w_i \end{aligned} \quad (3.60)$$

where $A_i = r_{i+1/2}/(r_i \Delta r^2)$ and $B_i = r_{i-1/2}/(r_i \Delta r^2)$. The bending pressure term P is very stiff and needs a careful treatment. In particular, the fourth order derivative requires a fourth order central difference scheme and therefore, P_i is expressed over a seven point stencil on the main grid such that

$$P_i = \alpha_i h_{i-3} + \beta_i h_{i-2} + \gamma_i h_{i-1} + \lambda_i h_i + \kappa_i h_{i+1} + \delta_i h_{i+2} + \varepsilon_i h_{i+3} \quad (3.61)$$

with

$$\begin{aligned} \alpha_i &= \frac{1}{24\Delta r^4} (-4 + 3p_3\Delta r) \\ \beta_i &= \frac{1}{24\Delta r^4} (48 - 24p_3\Delta r - 2p_2\Delta r^2 + 2p_1\Delta r^3) \\ \gamma_i &= \frac{1}{24\Delta r^4} (-156 + 39p_3\Delta r + 32p_2\Delta r^2 - 16p_1\Delta r^3) \\ \lambda_i &= \frac{1}{24\Delta r^4} (224 - 60p_2\Delta r^2) \\ \kappa_i &= \frac{1}{24\Delta r^4} (-156 - 39p_3\Delta r + 32p_2\Delta r^2 + 16p_1\Delta r^3) \\ \delta_i &= \frac{1}{24\Delta r^4} (48 + 24p_3\Delta r - 2p_2\Delta r^2 - 2p_1\Delta r^3) \\ \varepsilon_i &= \frac{1}{24\Delta r^4} (-4 - 3p_3\Delta r) \end{aligned}$$

and where $p_1 = 1/r_i^3$, $p_2 = 1/r_i^2$ and $p_3 = 2/r_i$. Finally, the term $\phi_{i-1/2}$ and $\phi_{i+1/2}$, which depend on the variable Θ_b , δ as well as different power of h , are evaluated in $i - 1/2$ and $i + 1/2$ respectively. Different choices for the value of the variable at the mid-cell grid point do not show any significant difference and a simple average is taken such that the variable $u_{i+1/2}$ is taken as $0.5(u_i + u_{i+1})$.

Expression of the jacobian J_f^h

The discretized function f_i can be break down in three part, the gravitational part f_i^g which is expressed in term of the value of h on three grid points $\{i - 1, i, i + 1\}$, the bending part f_i^b which is expressed in term of the value of h on nine grid points $\{i - 4, i - 3, \dots, i + 3, i + 4\}$ and the injection term which depends only on the grid point i such that

$$f_i = f_i^g + f_i^b + w_i \quad (3.62)$$

Therefore, the jacobian is nona-diagonal and its coefficient J_{il} are

$$J_{il} = \begin{cases} \frac{\partial f_i^b}{\partial h_l} & l = \{i - 4, i - 3, i - 2, i + 2, i + 3, i + 4\} \\ \frac{\partial f_i^g}{\partial h_l} + \frac{\partial f_i^b}{\partial h_l} & l = \{i - 1, i, i + 1\} \\ 0 & \text{otherwise} \end{cases} \quad (3.63)$$

The different terms can be easily derived from (3.60) and (3.61) with just slight adjustment coming from the boundary conditions.

Boundary condition

We begin with $h_i = h_f$ for $i = 1, \dots, M$. Since the flow is symmetric in $r = 0$, we require that

$$\left. \frac{\partial h}{\partial r} \right|_{r=0} = \left. \frac{\partial P}{\partial r} \right|_{r=0} = 0 \quad (3.64)$$

and therefore for $i = 1$, we have

$$\begin{aligned} f_i &= A_1 \phi_{i+1/2} (h_{i+1} - h_i) \\ &+ A_i \phi_{i+1/2} (P_{i+1} - P_i) \\ &+ w_i \end{aligned} \quad (3.65)$$

The expression of the bending pressure, evaluated over a 7 point stencils, is problematic close to the boundary and reflection formulae will be used in order to accommodate the boundary conditions [Patankar \(1980\)](#). In particular, we have $h_0 = h_1$, $h_{-1} = h_2$ and $h_{-2} = h_3$. Similarly, boundary condition at the

end of the mesh is accounted by using a grid much larger than the flow itself and requiring

$$\left. \frac{\partial h}{\partial r} \right|_{r=r_M} = \left. \frac{\partial P}{\partial r} \right|_{r=r_M} = 0 \quad (3.66)$$

which gives for $i = M$

$$\begin{aligned} f_i &= B_i \phi_{i-1/2} (h_i - h_{i-1}) \\ &+ B_i \phi_{i-1/2} (P_i - P_{i-1}) \\ &+ w_i \end{aligned} \quad (3.67)$$

with $h_{i>=M} = h_f$.

Newton-Rhapsod method

The Newton-Rhapsod method reads

$$(I - \Delta t J_f^h(h_k^*)) \delta h_k^n = h^n - h_k^* + \Delta t f(h_k^*) \quad (3.68)$$

where the k refers to the k iterations, I is a $M \times M$ diagonal matrix and $J_f^h(h^*)$ is a $M \times M$ nona-diagonal matrix. This system of linear equations can be solved using a nona-diagonal algorithm. At the first iteration, we use $h_1^* = h^n$ as a first guess and then we iterate using $h_k^* = h^n + \delta h_{k-1}^n$ as a new guess for each iterations until δh_k^n becomes sufficiently small. In particular, we require that

$$\delta h_k^n / h_k^* < \varepsilon \quad (3.69)$$

with $\varepsilon = 10^{-4}$.

3.3.3 Heat equation

The heat equation (3.46) is written as

$$\frac{\partial \xi}{\partial t} - g(h, \xi) = 0 \quad (3.70)$$

with

$$g = \frac{1}{r} \frac{\partial}{\partial r} (r \Gamma \xi) + \frac{1}{r} \frac{\partial}{\partial r} (r \Sigma) + 2Pe^{-1} St_m \frac{(\Theta_b - \Theta_s)}{\delta} \quad (3.71)$$

$$\Gamma = -\bar{u} \quad (3.72)$$

Spatial discretization of g

As for the thickness equation, the spatial discretization is obtained using a central difference scheme over a sub-grid shifted by $0.5\Delta r$ from the main grid. Therefore, we have

$$g_i = (C_i \Gamma_{i+1/2} \xi_{i+1/2} - D_i \Gamma_{i-1/2} \xi_{i-1/2}) \quad (3.73)$$

$$+ (C_i \Sigma_{i+1/2} - D_i \Sigma_{i-1/2}) \quad (3.74)$$

$$+ 2Pe^{-1} St_m \frac{\Theta_{b,i} - \Theta_{s,i}}{\delta_i} \quad (3.75)$$

with $C_i = r_{i+1/2}/(r_i \Delta r)$ and $D_i = r_{i-1/2}/(r_i \Delta r)$. We use the average between the grid point i and $i-1$ (resp. $i+1$) to evaluate the quantity in Γ and Σ at $i-1/2$ (resp. $i+1/2$). In addition, we use a classical upwind scheme to handle ξ at the mid grid point which requires

$$\xi_{i+1/2} = \xi_i \quad (3.76)$$

$$\xi_{i-1/2} = \xi_{i-1} \quad (3.77)$$

Expression of the Jacobian J_g^ξ

The expression of the Jacobian is much straightforward in that case and its coefficient J_{il} are

$$J_{il} = \begin{cases} -D_i \Gamma_{i-1/2} & l = i-1 \\ C_i \Gamma_{i+1/2} & l = i \\ 0 & \text{otherwise} \end{cases} \quad (3.78)$$

with only slight adjustment coming from the boundary conditions.

Boundary conditions

We consider $\Theta_b = 1$ and $\delta = 10^{-4}$ in the film at $t = 0$. In this way, we ensure that the average temperature across the film at $t = 0$ is close to 1. By construction, $D_1 = 0$ and therefore, for $i = 1$ we have

$$g_i = C_i \Gamma_{i+1/2} \xi_i + C_i \Sigma_{i+1/2} + 2Pe^{-1} St_m \frac{\Theta_{b,i} - \Theta_{s,i}}{\delta_i} \quad (3.79)$$

For $i = M$, we consider that $\Gamma_{i+1/2} = \Gamma_i$ and $\Sigma_{i-1/2} = \Sigma_i$. However, the choice for the boundary condition at the border of the grid $i = M$ is not important as we solve the problem over a grid much larger than the flow itself.

Newton-Rhapsod method

The Newton-Rhapsod method reads

$$(I - \Delta t J_g^\xi(\xi_k^*)) \delta \xi_k^n = \xi^n - \xi_k^* + \Delta t f(\xi_k^*) \quad (3.80)$$

where the k refers to the k iterations, I is a $M \times M$ diagonal matrix and $J_f^h(\xi^*)$ is a $M \times M$ tri-diagonal matrix. This system of linear equations can be solved using a tri-diagonal algorithm. As for the thickness equation, at the first iteration, we use $\xi_1^* = \xi^n$ as a first guess and then we iterate using $\xi_k^* = \xi^n + \delta \xi_{k-1}^n$ as a new guess for each iterations until $\delta \xi_k^n$ becomes sufficiently small. In particular, we require that

$$\delta \xi_k^n / \xi_k^* < \varepsilon \quad (3.81)$$

with $\varepsilon = 10^{-4}$. In addition, at each iteration the quantity $\Theta_{s,k}^*$, $\Theta_{b,k}^*$ and δ_k^* , that are needed to evaluate Γ and Σ , are derived from the value of ξ_k^* using (??), (??) and (??) respectively.

3.4 Evolution in the bending regime

We first concentrate on the case in which only bending contributes to the dynamics pressure. The governing equations are thus (3.45) and (3.46) where $P = \nabla_r^4 h$.

3.4.1 Thermal structure for an isoviscous flow, effect of Pe

The current cools by conduction and thermal boundary layers form at the contact with the surrounding medium. These boundary layers first connect at the tip of the flow, where the small thickness induces an important cooling (Figure 3.3). A region of cold fluid forms at the front.

As the current thickens with time, a balance between advection and diffusion of heat is never reached in the interior of the current. The hot thermal anomaly grows in extent with time but it extends slower than the current itself and the cold fluid region at the tip grows even faster. For instance, for $Pe = 100$, while the region of cold fluid extends over about 10% of the current at $t = 0.5$, it extends over about 20% at $t = 10$ (Figure 3.3). The smaller Pe , the more important the conductive cooling and the larger the cold fluid region is (Figure 3.4 and 3.5). For instance, at $t = 10$, while the cold fluid region extends over about 20% of the current for $Pe = 100$, it extends over more than 70% for $Pe = 1$ (Figure 3.5).

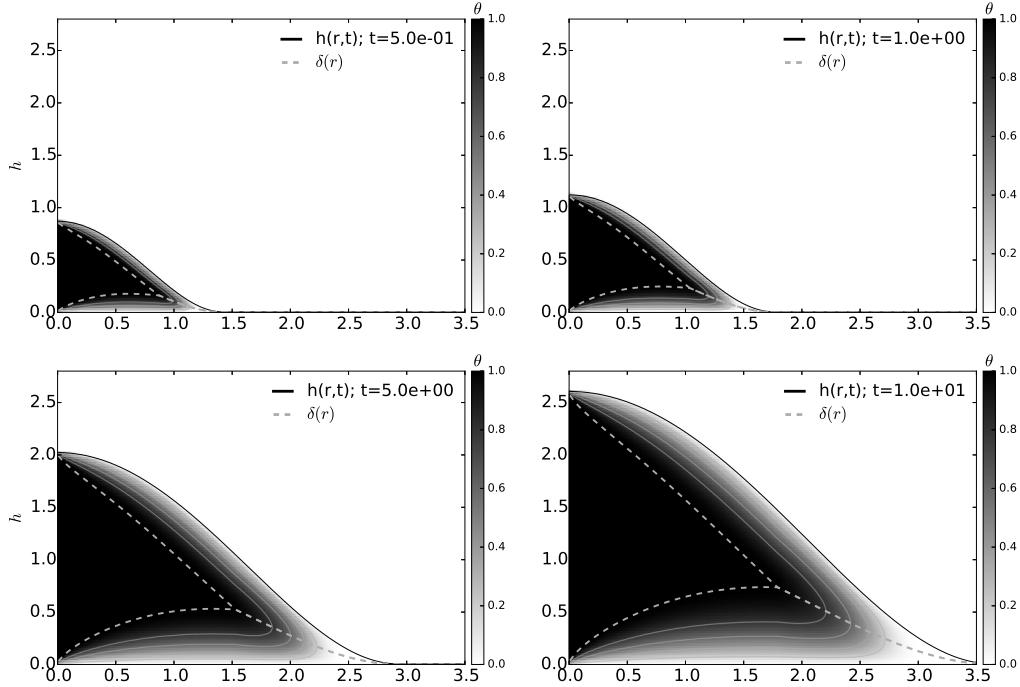


Figure 3.3: Snapshots of the flow thermal structure $\theta(r, z, t)$ at different times indicated on the plot. Dashed lines represent the thermal boundary layers. Solid grey lines are isotherms for $\theta = 0.2, 0.4, 0.6$ and 0.8 . Here, $\nu = 1$, $Pe = 100$, $St_m = 1$.

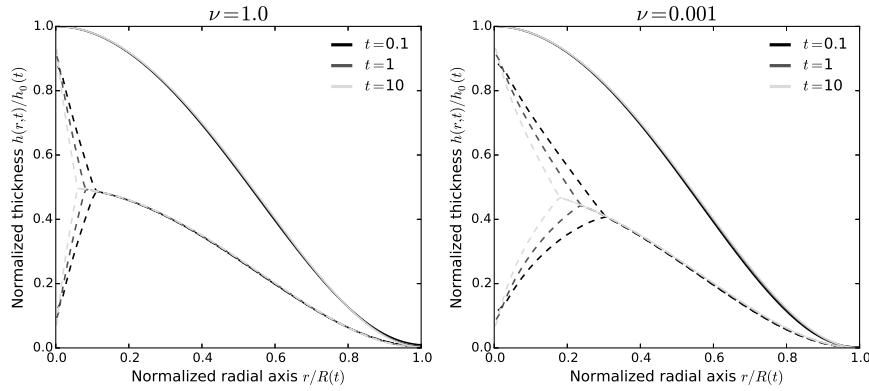


Figure 3.4: Left: thickness normalized by the thickness at the center $h(r, t)/h_0(t)$ versus radial axis normalized by the current radius $r/R(t)$ at different times indicated on the plot for $Pe = 1$ and $\nu = 1.0$. Solid-lines represent the thickness profiles. Dashed-lines represent the thermal boundary layers. Right: Same plot but for $\nu = 10^{-3}$.

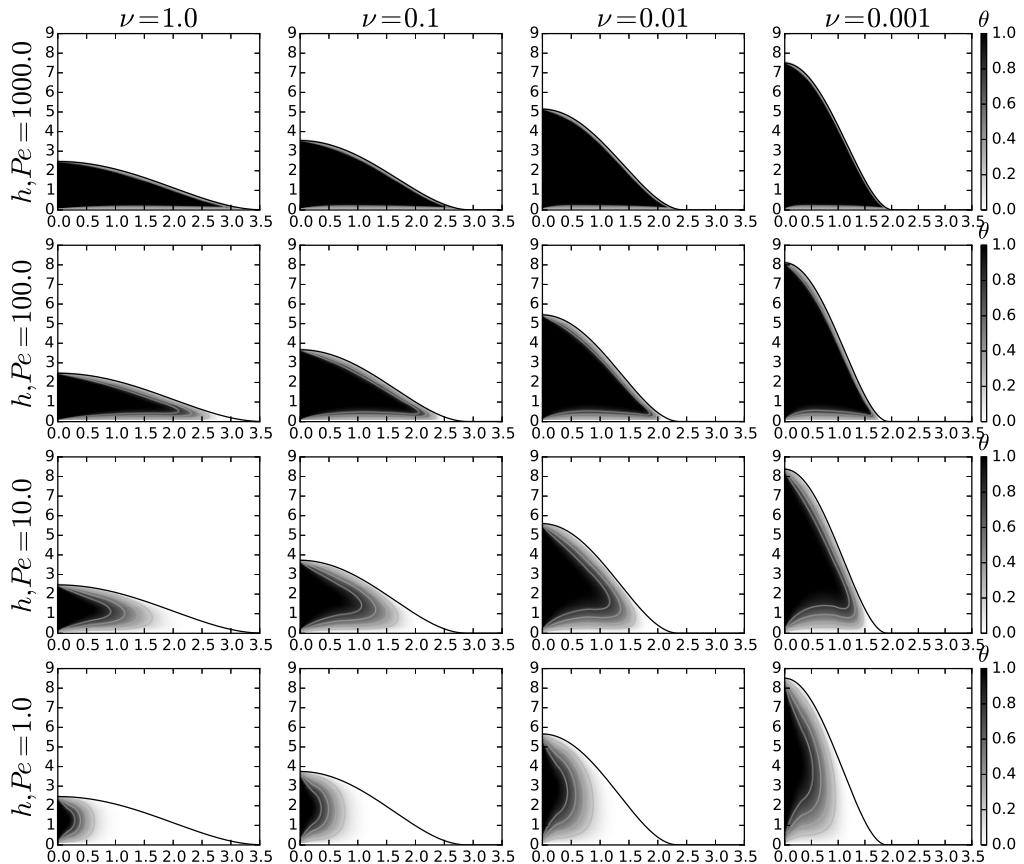


Figure 3.5: Snapshots of the flow thermal structure $\theta(r, z, t)$ for different set (ν, Pe) with $\nu = 1, 0.1, 0.01$ and 0.001 and $Pe = 1, 10, 100$ and 1000 at $t = 10$. While Pe controls the thermal structure of the flow, it has only a small influence on the flow aspect ratio which is controlled by ν .

3.4.2 Thickness and temperature profile, effect of ν

When accounting for the temperature dependence of the viscosity, the region of cold fluid at the tip is marked by a higher viscosity and enhances flow thickening at the expense of spreading. The larger the viscosity contrast, the larger the aspect ratio h_0/R (Figure 3.5). For instance, for the same value of $Pe = 1$, while the aspect ratio is 0.7 for $\nu = 1$ at $t = 10$, it is 4.2 for the same time and $\nu = 10^{-3}$ (Figure 3.5). Nevertheless, the shape of the flow remains essentially self-similar (3.49) and cannot be differentiated from the shape of an isoviscous current if the thickness and the radial coordinates are rescaled by the thickness at the center $h_0(t)$ and radius $R(t)$ (Figure 3.4).

The flow thermal structure is similar to the isoviscous case (Figure 3.5), the thermal anomaly rapidly detaches from the tip of the current and a region of cold fluid develops at the front where the heat loss is largest. However, the important thickening induced by the viscosity increase limits heat loss to the surrounding. The larger the viscosity contrast ν , the more important the thickening and the larger the thermal anomaly at a given time. For instance, for $Pe = 1$, while the thermal anomaly extends over about 30% of the flow for $\nu = 1$ at $t = 10$, it extends over more than 50% for $\nu = 10^{-3}$ (Figure 3.5).

As expected, a larger Peclet number leads to a larger thermal anomaly (Figure 3.5). However, although different Peclet numbers cause very different thermal structures, the influence of the Peclet number on the flow morphology is small, much smaller than the effect of the viscosity contrast ν (Figure 3.5). For instance, for $\nu = 10^{-3}$ at $t = 10$, the thermal anomaly is still attached to the tip of the current for $Pe = 1000$ whereas it makes about 50% of the current for $Pe = 1$; but, the thickness h_0 and the radius R in both cases differ only by a few percents (Figure 3.5). This suggests that the spreading of the flow is not controlled by the mean temperature or average viscosity of the flow.

3.4.3 Evolution of the thickness and the radius

The dynamics show three different spreading phases. The thickness as well as the radius first follow the isoviscous scaling laws for a hot viscosity current $h_0 \propto t^{8/22}$ (3.51) and $R \propto t^{7/22}$ (3.52) (Figure 3.6). In the second phase, thickening occurs at the expense of spreading because the thermal anomaly has detached from the current radius and the viscous cold fluid region at the front slows down the spreading. Finally, the dynamics enters a third phase where the thickness and radius follow the scaling laws for the spreading of an isoviscous current characterized by a dimensionless cold viscosity $1/\nu$. These scaling laws are obtained from (3.51) and (3.52) by rescaling the characteristic

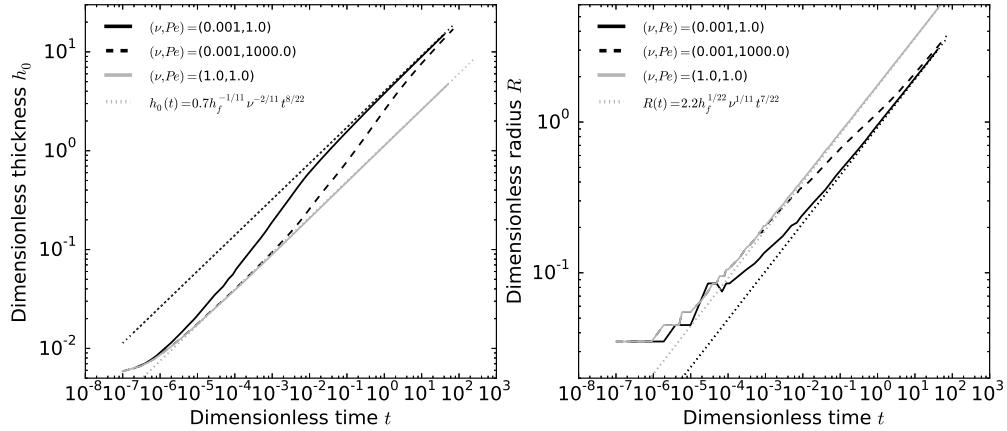


Figure 3.6: Left: Dimensionless thickness at the center h_0 versus dimensionless time t for different sets (ν, Pe) indicated on the plot. Dotted-lines: scaling laws $h_0 = 0.7h_f^{-1/11}\nu^{-2/11}t^{8/22}$ for $\nu = 1.0$ and 0.001 . Right: Dimensionless radius R versus dimensionless time t for the same sets (ν, Pe) . Dotted-lines: the scaling laws $R = 2.2h_f^{1/22}\nu^{1/11}t^{7/22}$ for $\nu = 1.0$ and 0.001 .

thickness and time by $\nu^{1/4}$ and read

$$h_0 = 0.7\nu^{-2/11}h_f^{-2/22}t^{8/22} \quad (3.82)$$

$$R = 2.2\nu^{1/11}h_f^{1/22}t^{7/22}. \quad (3.83)$$

The dependence on the viscosity contrast ν indeed fits very well the third phase of the flow observed in the numerical simulations (Figure 3.6). These results suggest that the effective viscosity η_e that governs the flow dynamics is first close to the viscosity of the hot fluid; it rapidly increases in the second phase to asymptotically tend to the one of the cold fluid in the third phase.

The time the flow spends in each phase depends on the Peclet number Pe . For instance, for $\nu = 10^{-3}$, while the current leaves the first phase at $t \sim 10^{-6}$ for $Pe = 1.0$, this transition happens only after $t \sim 10^{-2}$ for $Pe = 10^3$ (Figure 3.6). The larger the Peclet number, the less efficient the cooling, and thus the longer the flow remains in the first phase and the later it reaches the third phase.

3.4.4 Characterization of the thermal anomaly

Following *Garel et al. (2012)*, we quantify the size of the thermal anomaly through a critical thermal radius $R_c(t)$ where the temperature at the center of the flow Θ_b is 1% of the injection temperature, i.e. $\Theta_b(r = 0) - \Theta_b(r = R_c) = 0.99$.

The thermal anomaly is first advected at the same velocity as the current itself, i.e. $R(t) = R_c(t)$ (Figure 3.7 left). After a time that depends on Pe and ν , the thermal anomaly detaches from the tip and $R(t) - R_c(t)$ increases with time (Figure 3.7).

In the bending regime, the interior pressure is constant and the thickness profile $h(r)$ is given by (3.49) (Figure 3.4). The size of the thermal anomaly $R_c(t)$ is given by the radius where advection of heat is equal to heat loss

$$\frac{d}{dt}(\theta(r = R_c, t)) \propto Pe^{-1} \frac{\partial^2}{\partial z^2}(\theta(r = R_c, t)). \quad (3.84)$$

Assuming that, at the edge of the thermal anomaly, θ is constant and close to Θ_b , i.e. $\theta \approx \Theta_b$, we obtain by integrating (3.84) over h and using (3.21)

$$\begin{aligned} \frac{d}{dt} \left(\int_0^h \theta dz \right) &\propto Pe^{-1} \frac{\Theta_b}{h} \\ \Theta_b \frac{dh}{dt} &\propto Pe^{-1} \frac{\Theta_b}{h} \\ \frac{dh}{dt} &\propto \frac{Pe^{-1}}{h}. \end{aligned} \quad (3.85)$$

Using the thickness profile (3.49), (3.85) becomes

$$\alpha^2 \left(1 + \frac{R_c}{R}\right)^2 \frac{\partial h_0}{\partial t} + \frac{4h_0 R_c^2}{R^3} \frac{\partial R}{\partial t} \alpha \left(1 + \frac{R_c}{R}\right) = \frac{Pe^{-1}}{\alpha^2 \left(1 + \frac{R_c}{R}\right)^2 h_0}$$

where we introduce $\alpha(t) = (R(t) - R_c(t))/R(t)$, i.e. the normalized region beyond $r = R_c(t)$. In the limit $\alpha \ll 1$, i.e. $R_c/R \sim 1$ and discarding higher-order terms, we finally get

$$\alpha^3 \propto \frac{Pe^{-1}}{h_0^2(t)} \frac{R}{\frac{\partial R}{\partial t}}. \quad (3.86)$$

Substituting the thickness $h_0(t)$ and the radius $R(t)$ by their respective scaling laws (3.82) and (3.83), the relative size of the normalized cold front region α reads

$$\alpha(t) \propto Pe^{-1/3} \nu^{4/33} h_f^{2/33} t^{1/11}. \quad (3.87)$$

which is equivalent to

$$R(t) - R_c(t) = 2.1 Pe^{-1/3} \nu^{7/33} h_f^{7/66} t^{9/22} \quad (3.88)$$

where the numerically prefactor, which depends on the definition of the thermal anomaly, has been chosen to fit the simulations.

The predicted scaling law for the extent of the cold fluid region (3.88) indeed closely fits the numerical simulations for $\nu < 1$ and the different Peclet numbers (Figure 3.7). For $\nu = 1$ and $Pe = 1$, the condition $R - R_c \ll R$ is no more respected for $t > 0.1$, the thermal anomaly is much smaller than the flow itself and the scaling law (3.88) is no more applicable as expected.

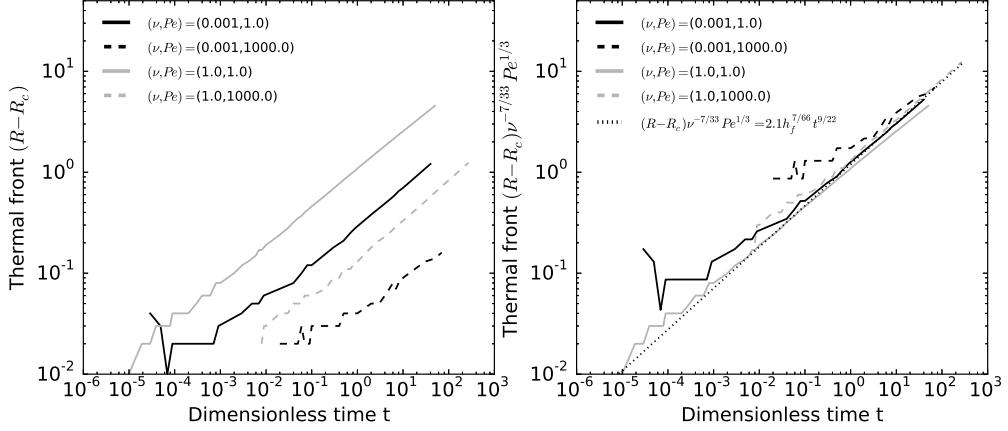


Figure 3.7: Left: Extent of the cold fluid region $R(t) - R_c(t)$ versus dimensionless time for different combinations (ν, Pe) indicated on the plot. Right: Same plot but where we rescale the extent of the cold fluid region by $Pe^{-1/3} \nu^{7/33}$. Dotted-line: scaling law $(R(t) - R_c(t))Pe^{1/3}\nu^{-7/33} = 2.1h_f^{7/66}t^{9/22}$.

3.4.5 Effective viscosity of the current

We use the predicted scaling law for the thickness $h_0(t)$ (3.82) to infer the time evolution of the effective viscosity $\eta_e(t)$. Indeed, substituting ν by $\eta_h/\eta_e(t)$ in (3.82) and inverting for $\eta_e(t)/\eta_h$, we get

$$\eta_e(t)/\eta_h = \left(\frac{h_0(t)t^{-8/22}}{0.7h_f^{-2/22}} \right)^{11/2} \quad (3.89)$$

where $h_0(t)$ is given by the simulation.

As suggested by the results of section 3.4.3, the effective viscosity is first close to the hot viscosity η_h , i.e. $\eta_e/\eta_h \sim 1$. It rapidly increases in the second phase of propagation and finally tends to the cold viscosity η_c in the third phase, i.e. $\eta_e/\eta_h \sim 1/\nu$ (Figure 3.8 top). The spreading in the isoviscous case is controlled by the propagation of a peeling by bending wave at the tip of the current (Lister *et al.*, 2013). In agreement, the behavior of the effective viscosity has to be linked with the rapid cooling of the front. To test this hypothesis, we calculate the average viscosity $\eta_f(t)$ over a fixed front region of size L in between $R(t) - L$ and $R(t)$ such that

$$\eta_f/\eta_h = \frac{1}{V_f} \int_{R-L}^R \int_0^h r\eta(\theta) dr dz \quad (3.90)$$

where $V_f(t)$ is the volume of this region. The numerical evaluation of $\eta_f(t)$ for a constant size $L \sim 0.1$ gives a good agreement with the evolution of

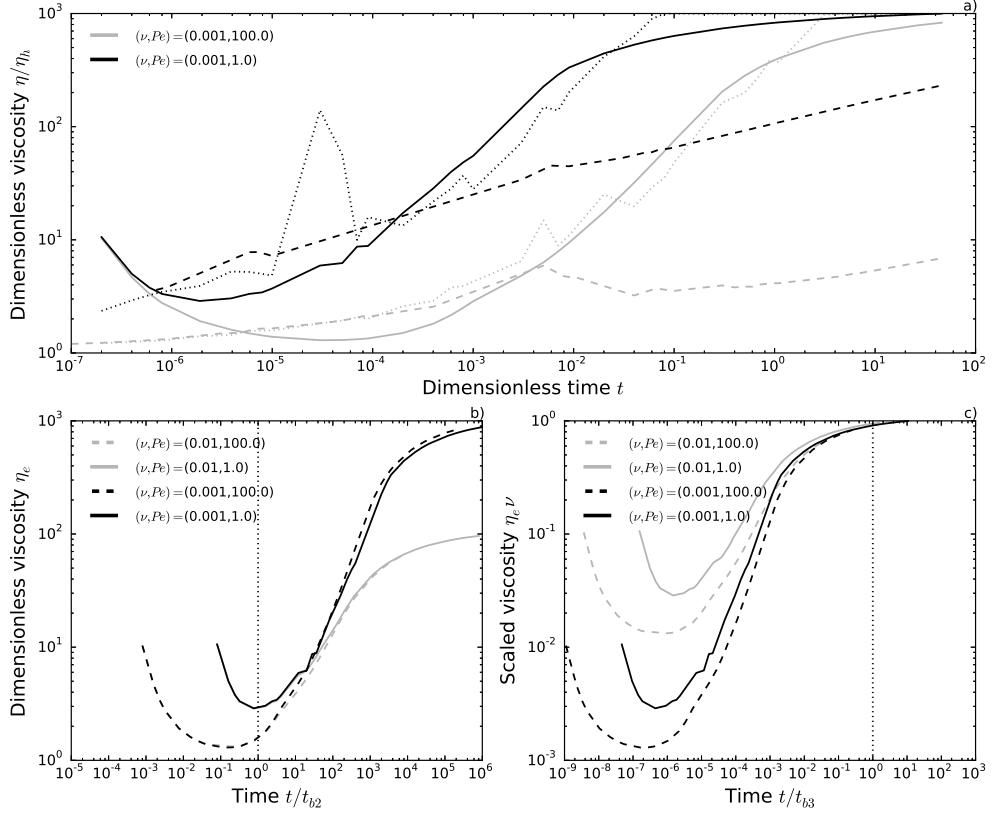


Figure 3.8: Top: Dimensionless viscosity $\eta(t)/\eta_h$ versus dimensionless time t for different combinations (ν, Pe) indicated on the plot. Solid lines: effective viscosity η_e/η_h defined by (3.89). Dashed-lines: average flow viscosity defined by $\eta_a(t)/\eta_h = \frac{1}{V(t)} \int_0^{R(t)} \int_0^{h(r,t)} r\eta(\theta)drdz$ where $V(t)$ is the current volume. Dotted-lines: average front viscosity η_f/η_h defined by (3.90). Bottom left: dimensionless effective viscosity η_e versus time where the time has been rescaled by the time for the flow to enter the second phase t_{b2} . Bottom right: Same as left but where the time has been rescaled by the time for the flow to enter the third phase t_{b3} .

the effective viscosity η_e for the second phase of propagation (Figure 3.8). Therefore, the effective viscosity, and thus the different phases of propagation, are controlled by the average viscosity of a small region at the front of the current.

At the initiation of the flow, the pre-wetted film is composed by fluid at the injection temperature, the thermal anomaly is attached to the front and the current spreads with a hot viscosity η_h . As soon as the film has cooled, the thermal anomaly detaches from the tip of the current and the effective viscosity increases. The time t_{b2} the current enters this second phase of the flow thus scales as the time to cool the pre-wetted film thickness by conduction, i.e. $t_{b2} = 0.1Pe h_f^2$ where the numerical prefactor has been matched to the simulations. Indeed, when rescaling the time of the simulations by t_{b2} , the different combinations (ν, Pe) enter the second phase simultaneously (Figure 3.8, bottom left). Then, the size of the cold fluid region at the front increases, the effective viscosity increases, and when $R(t) - R_c(t)$ becomes larger than ~ 0.1 , the current behaves as an isoviscous current with cold viscosity η_c . Therefore, the time t_{b3} for the flow to enter this third phase scales as the time for the cold fluid region, whose size is given by (3.88), to be larger than ~ 0.1 . In particular, we define the time t_{b3} as the time for the effective to reach 90% of its maximum value η_c . Inverting (3.88) and matching the numerical prefactor to the simulation thus gives $t_{b3} \sim 0.01Pe^{22/27}\nu^{-14/27}h_f^{-7/27}$. Indeed, when rescaling the time of the simulations by t_{b3} , the different combinations (ν, Pe) enter the third phase simultaneously (Figure 3.8, bottom right).

3.4.6 Note on the effect of crystallization

Here, we examine the effect of crystallization on the flow dynamics and use values of $St_m < 1$. Crystallization induces a release of latent heat in the fluid, increasing the amount of available energy at a given time. When $St_m < 1$, the tip of the current remains hot for a longer time and the flow transitions to the second phase later than in the case where $St_m = 1$ (Figure 3.9). As the crystallization acts only to reduce the cooling term by a factor St_m in (3.46), one can easily rewrite (3.88) to account for the effect of crystallization on the size of the cold fluid region

$$R(t) - R_c(t) = 2.1Pe^{-1/3}St_m^{1/3}\nu^{7/33}h_f^{7/66}t^{9/22}. \quad (3.91)$$

Indeed, the dependence with the dimensionless number St_m is well described by the scaling law (3.91) (Figure 3.10). Accordingly, the time t_{b2} and t_{b3} for the current to enter the second and third phase of the flow respectively are

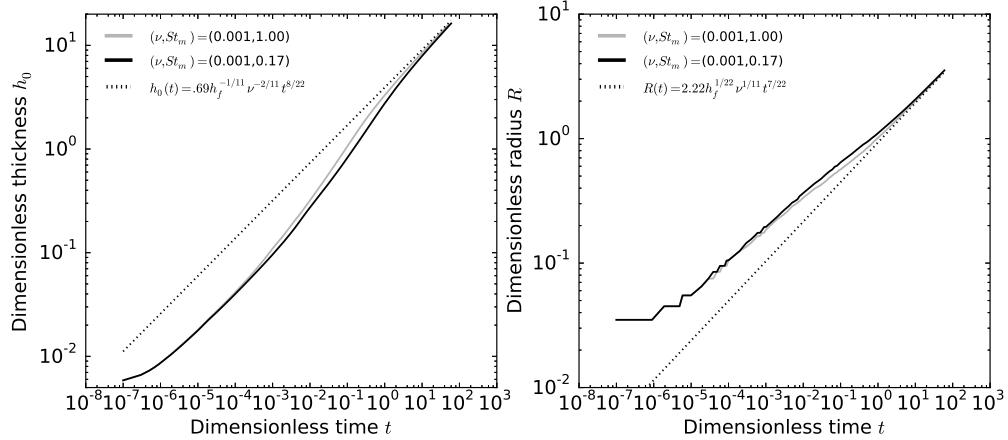


Figure 3.9: Left: Dimensionless thickness at the center h_0 versus dimensionless time t for different values of St_m indicated on the plot, $\nu = 0.001$ and $Pe = 10.0$. Dotted-line: scaling law $h_0 = 0.7h_f^{-1/11}\nu^{-2/11}t^{8/22}$ for $\nu = 0.001$. Right: Dimensionless radius R versus dimensionless time t for the same combinations of dimensionless numbers. Dotted lines: scaling law $R = 2.2h_f^{1/22}\nu^{1/11}t^{7/22}$ for $\nu = 0.001$.

delayed and when accounting for crystallization read

$$t_{b2} \sim 0.1PeSt_m^{-1}h_f^2 \quad (3.92)$$

$$t_{b3} \sim 10^{-2}St_m^{-22/27}Pe^{22/27}\nu^{-14/27}h_f^{-7/27} \quad (3.93)$$

3.5 Evolution in the gravity current regime

To study the late time behavior, we concentrate on the case where only the weight of the fluid contributes to the pressure. The governing equation are thus (3.45) and (3.46) where $P = h$. We follow the same development as in Section 3.4. In particular, we first describe the thermal structure for an isoviscous flow, i.e. $\nu = 1$, then we study the effect of the temperature-dependent viscosity on the current dynamics without crystallization, i.e. $\nu < 1$ and $St_m = 1$. Finally, we look at the effect of crystallization by setting $St_m < 1$.

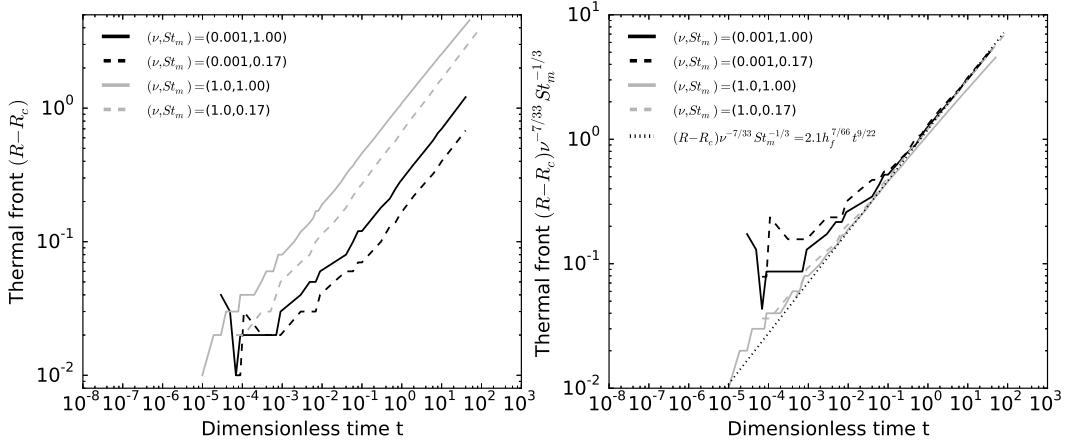


Figure 3.10: Left: Extent of the cold fluid region $R(t) - R_c(t)$ versus dimensionless time for different combinations (ν, St_m) indicated on the plot and $Pe = 1$. Right: Same plot but where we have rescaled the extent of the cold fluid region by $St_m^{-1/3} \nu^{7/33}$. Dotted-line: scaling law $(R(t) - R_c(t)) St_m^{-1/3} \nu^{-7/33} = 2.1 h_f^{7/66} t^{9/22}$.

3.5.1 Thermal structure for an isoviscous flow, effect of Pe

As in the bending regime, the bulk of the fluid first expands at the injection temperature and $R_c \sim R$. As the bottom and the top cool by conduction, thermal boundary layers form at the contact with the surrounding medium and connect at the tip of the current. However, in the gravity current regime, the thickness of the current tends to a constant. Therefore, conduction in the surrounding medium rapidly balances the input of heat at the center and when the thermal anomaly detaches from the tip of the current, its extent reaches a steady state profile (Figure 3.11).

The radius of the steady state thermal anomaly R_c depends on Pe . In particular, the larger the number Pe , the larger the radius R_c is. For instance, for $\nu = 1$, while the thermal anomaly R_c is less than 1 in the steady state regime for $Pe = 1$, it is about 12 for $Pe = 10^3$ (Figure 3.12).

3.5.2 Thickness and temperature profile, effect of ν

For a current with a viscosity that depends on temperature, as soon as the thermal anomaly detaches from the current radius, the cold fluid at the front tends to slow down the spreading and enhance the thickening of the flow

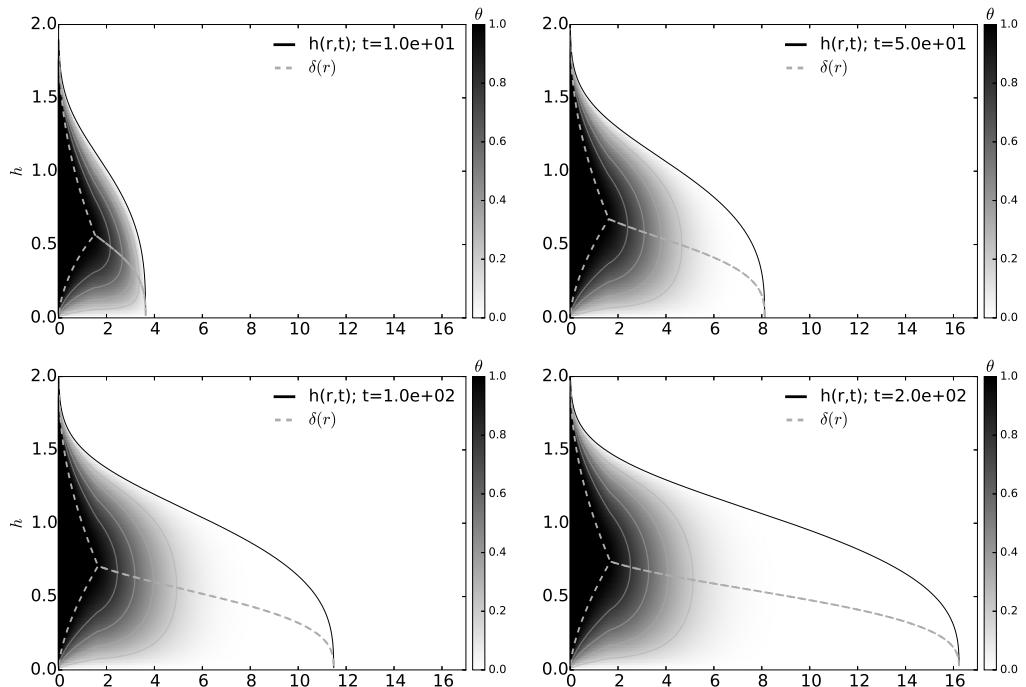


Figure 3.11: Snapshots of the flow thermal structure $\theta(r, z, t)$ at different times indicated on the plot. Dashed lines: thermal boundary layers. Here, $\nu = 1$, $Pe = 100$ and $St_m = 1$.

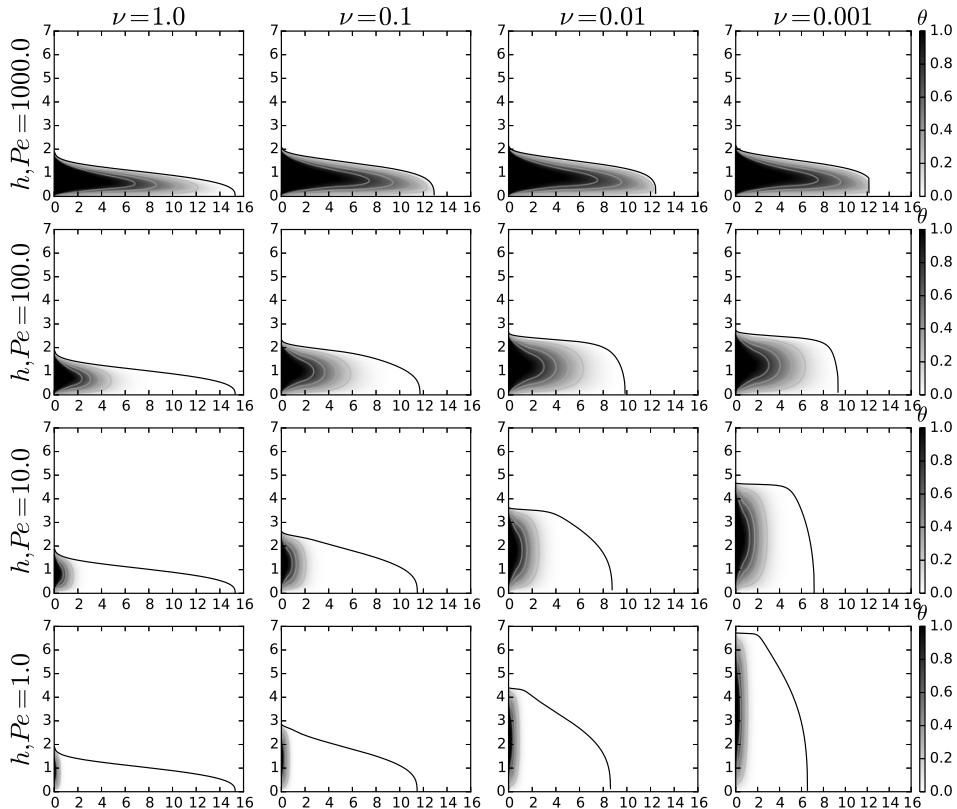


Figure 3.12: Snapshots of the flow thermal structure $\theta(r, z, t)$ for different sets (ν, Pe) with $\nu = 1, 0.1, 0.01$ and 0.001 and $Pe = 1, 10, 100$ and 1000 at $t = 200$.

(Figure 3.12). For instance, for $Pe = 1$, while the aspect ratio h_0/R is about 0.12 for $\nu = 1$ at $t = 200$, it is ~ 1 for $\nu = 10^{-3}$ (Figure 3.12). The shape of the current is not self-similar and the front steepens when the viscosity increases in comparison to the isoviscous case as in *Bercovici (1994)*. However, when the current becomes much larger than the thermal anomaly, the current side slumps to become less steep (Figure 3.12) and recovers a shape similar to the isoviscous flow with cold viscosity.

The thermal structure is similar to the isoviscous case. In particular, after a time that depends on Pe , the thermal anomaly reaches a steady-state profile (Figure 3.12). As in the bending regime, the thickening at the center limits heat loss to the surrounding for large values of the viscosity contrast ν . Therefore, the extent of the thermal anomaly in the steady-state is slightly larger for a larger viscosity contrast. For instance, for $Pe = 10$ at $t = 200$, while the thermal anomaly extends over less than 2 for $\nu = 1$, it reaches $Rc \sim 3$ for $\nu = 10^{-3}$.

The flow morphology is more sensitive to the Pe number in the gravity current regime than in the bending regime and different Pe lead to different current morphologies for a given ν (Figure 3.12). For instance, for $\nu = 10^{-3}$ at $t = 200$, the thermal anomaly is still attached to the tip of the current for $Pe = 10^3$ and the aspect ratio of the flow h_0/R is close to 0.15. In contrast, for $Pe = 1$, the thermal anomaly radius R_c is less than 30% of the current radius and the aspect ratio of the flow is much larger $h_0/R = 1.15$ (Figure 3.12).

3.5.3 Evolution of the thickness and the radius

As in the bending regime, the dynamics show three different spreading phases. The thickness as well as the radius first follow the isoviscous scaling laws for a given hot viscosity η_h , i.e. h_0 tends to a constant and $R \propto t^{1/2}$ (Figure 3.13). In a second phase, the thickness rapidly increases and the spreading slows down. Finally, the thickness and radius follow the isoviscous scaling laws but for a cold viscosity flow.

These dimensionless scaling laws read, as a function of ν

$$h_0 = 2.1\nu^{-1/4} \quad (3.94)$$

$$R(t) = 1.1\nu^{1/8}t^{1/2} \quad (3.95)$$

They perfectly matched our numerical simulations (Figure 3.13). Therefore, the effective viscosity η_e that controls the flow dynamics is first close to the viscosity of the hot fluid η_h ; it then rapidly increases to asymptotically tend to the viscosity of the cold fluid η_c in the third phase.

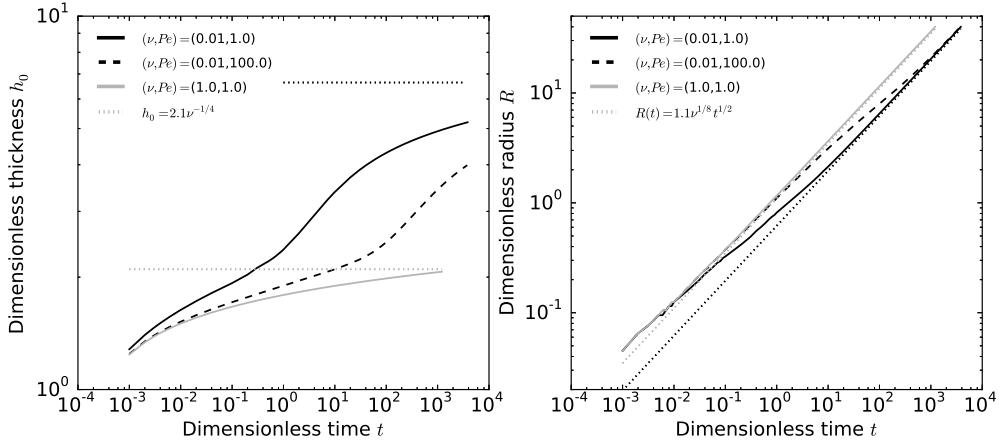


Figure 3.13: Left: Dimensionless thickness at the center h_0 versus dimensionless time t for different sets (ν, Pe) indicated on the plot. Dotted-lines represent the scaling laws $h_0 = 2.1\nu^{-1/4}$ for $\nu = 1.0$ and 10^{-2} . Right: Dimensionless radius R versus dimensionless time t for the same sets (ν, Pe) . Dotted-lines represent the scaling laws $R = 1.1\nu^{1/8}t^{1/2}$ for $\nu = 1.0$ and 10^{-2} .

As in the bending regime, the time the current spends in each phase depends on Pe (Figure 3.13). For instance, for $\nu = 10^{-2}$, while the current leaves the first phase at $t \sim 10^{-1}$ for $Pe = 1.0$, the transition occurs after $t \sim 10^1$ for $Pe = 10^2$. In general, the larger Pe , the longer the current remains in the first phase and the later is reached the third phase.

3.5.4 Characterization of the thermal anomaly

The thermal anomaly is first advected at the same velocity as the current itself, i.e. $R_c(t)/R(t) \sim 1$ (Figure 3.14 left). After a time that depends on Pe and ν , the thermal anomaly detaches from the front and reaches a steady-state profile (Figure 3.12 and 3.14).

We propose a simple thermal budget to predict the extent of the thermal anomaly in the steady-state regime. When the size of the thermal anomaly reaches a steady state, a balance between heat advection and diffusion in the surrounding medium in a dimensional form gives

$$\rho C_p U_0 \frac{\Delta T}{R_c} = \frac{8k\Delta T}{h_0^2} \quad (3.96)$$

where ΔT hold for a mean temperature contrast between the fluid and the surroundings advected at a mean velocity U_0 . For a gravity current, and by opposition to the bending regime, the thickness h_0 reaches a constant. Taking

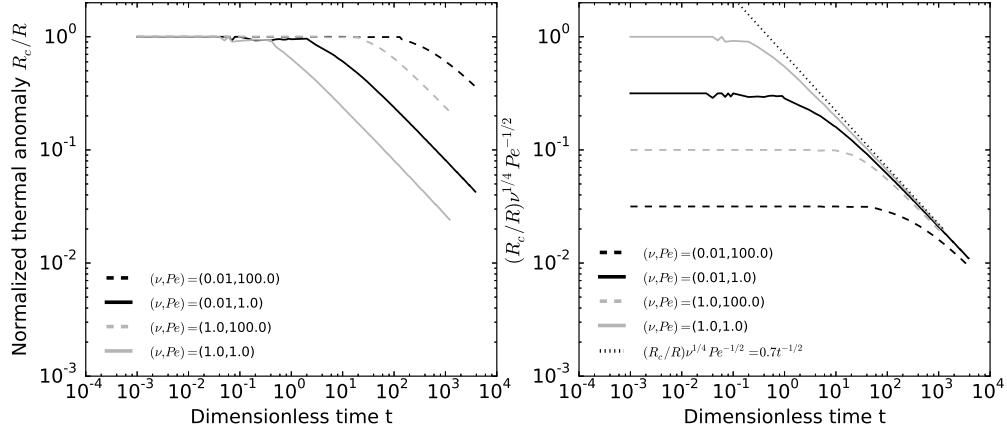


Figure 3.14: Left: Normalized thermal anomaly radius $R_c(t)/R(t)$ versus dimensionless time for different combinations (ν, Pe) indicated on the plot. Right: Same plot but where we rescale the normalized thermal anomaly radius $R_c(t)/R(t)$ by $Pe^{1/2}\nu^{-1/4}$.

U_0 as a horizontal redistribution of the injection rate, we write

$$U_0 = Q_0/(2\pi R_c h_0) \quad (3.97)$$

which gives

$$R_c = \frac{1}{4} \sqrt{\frac{h_0 Q_0}{\pi \kappa}} \quad (3.98)$$

By non-dimensionalizing (3.98), we finally get the expression for the thermal anomaly radius R_c in the steady state regime. In particular, we have $R_c \propto Pe^{1/2}\nu^{-1/8}$ and then

$$\frac{R_c}{R(t)} = 0.7 Pe^{1/2}\nu^{-1/4} t^{-1/2} \quad (3.99)$$

where we have used (3.95) and the numerical prefactor, which depends on the definition of the thermal anomaly, has been chosen to fit the simulations.

The analytical solution for the normalized thermal anomaly radius $R_c/R(t)$ (3.99) closely fits the numerical simulations (Figure 3.14). Indeed, when the thermal anomaly enters the steady state, the thermal anomaly radius remains constant and the normalized thermal anomaly radius $R_c(t)/R(t)$ evolves as the inverse of the current radius, i.e. as $t^{-1/2}$ (Figure 3.14). Furthermore, both the dependence with Pe and ν vanish when rescaling $R_c/R(t)$ by $Pe^{1/2}\nu^{-1/4}$ in the steady state regime (Figure 3.14, right).

3.5.5 Effective viscosity of the current

Repeating the same exercise as in section (3.4.5), we use the predicted scaling law for the radius $R(t)$ (3.95) to infer the effective viscosity $\eta_e(t)$ of the current

$$\eta_e(t)/\eta_h = \left(\frac{R(t)t^{-1/2}}{1.1} \right)^{-8}. \quad (3.100)$$

As expected, the effective viscosity in the gravity current regime represents the average viscosity of the current and the different phases of propagation reflect changes in the average viscosity of the flow (Figure 3.15).

At flow initiation, the thermal anomaly is advected at the same velocity as the current itself and the current spreads with hot viscosity η_h . When the thermal anomaly detaches from the tip and enters a steady state, η_e increases. The time t_{g2} to enter this second phase scales with the time to cool the current by conduction, i.e. $t_{g2} = 10^{-2}Pe$ where the numerical pre-factor has been matched to the simulations. Indeed, when rescaling the time by t_{g2} , the different combinations (ν, Pe) enter the second phase simultaneously (Figure 3.15, bottom left). Then, the size of the cold fluid region at the front increases, the effective viscosity increases and, when the current is large compared to the steady-state thermal anomaly radius, i.e. $R_c/R < 0.6$ the current behaves as an isoviscous current with cold viscosity η_c . Therefore, the time t_{g3} for the flow to enter this third phase scales as the time for the normalized thermal anomaly radius to be smaller than ~ 0.6 . As in the bending regime, we define the time t_{g3} as the time for the effective to reach 90% of its maximum value η_c . Inverting (3.88) and matching the numerical prefactor to the simulation thus gives $t_{g3} = 5Pe\nu^{-1/2}$. In agreement, when rescaling the time of the simulations by t_{g3} , the different combinations (ν, Pe) enter the third phase simultaneously (Figure 3.15, bottom right).

3.5.6 Note on the effect of crystallization

As in the bending regime, crystallization induces a release of latent heat to the fluid, increasing the amount of available energy at a given time. As a result, when $St_m < 1$, the current is hotter on average and it transitions to the second phase later than in the case where $St_m = 1$ (Figure 3.16). As in section (3.4.6), one can easily rewrite (3.99) to account for the effect of crystallization on the thermal anomaly evolution

$$\frac{R_c}{R(t)} = 0.7St_m^{-1/2}Pe^{1/2}\nu^{-1/4}t^{-1/2} \quad (3.101)$$

Indeed, the dependence with the dimensionless number St_m is well described by the scaling law (3.101) (Figure 3.17). Accordingly, the time t_{g2} and t_{g3} for

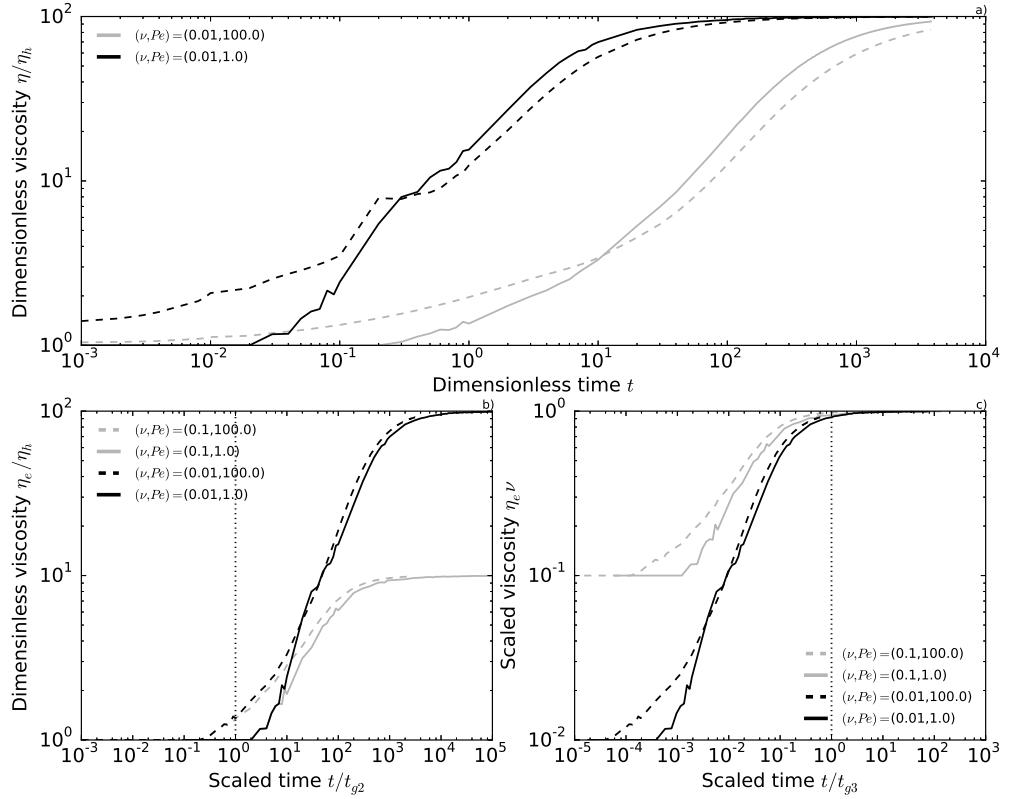


Figure 3.15: Top: Dimensionless viscosity $\eta(t)/\eta_h$ versus dimensionless time t for different combinations (ν, Pe) indicated on the plot. Solid lines: effective viscosity η_e/η_h defined by (3.89). Dashed-lines: average flow viscosity defined by $\overline{\eta_a(t)}/\eta_h = \frac{1}{V(t)} \int_0^{R(t)} \int_0^{h(r,t)} r\eta(\theta)drdz$ where $V(t)$ is the current volume. Bottom left: dimensionless effective viscosity η_e versus time where the time has been rescaled by the time t_{g2} (3.102). Bottom right: Same as left but where the time has been rescaled by t_{g3} (3.103).

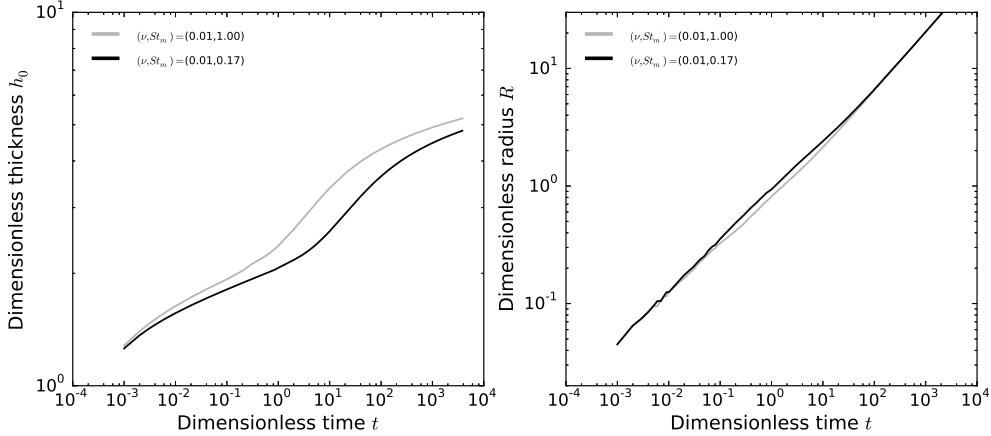


Figure 3.16: Left: Dimensionless thickness at the center h_0 versus dimensionless time t for different sets (ν, St_m) indicated on the plot and $Pe = 1$. Right: Dimensionless radius R versus dimensionless time t for the same sets (ν, St_m) and $Pe = 1$.

the current to enter the second and third phase of the flow are both delayed and finally read

$$t_{g2} \sim 10^{-2} Pe St_m^{-1} \quad (3.102)$$

$$t_{g3} \sim 5 Pe St_m^{-1} \nu^{-1/2} \quad (3.103)$$

3.6 Different evolutions with bending and gravity

For an isoviscous flow with $h_f \ll h \ll d_c$, the flow passes through three asymptotic dynamical regimes ([Michaut, 2011](#); [Lister et al., 2013](#)). When the radius R is much smaller than a critical radius $R_c \sim 4$, the interior solution is bell-shaped and peeling by bending controls propagation. In contrast, when $R \gg R_c$, bending stresses can be neglected almost everywhere and the flow enters a gravity current regime. In between, [Lister et al. \(2013\)](#) also describe a short intermediate regime where the peeling by bending continues to control the propagation but where the flow shows an interior flat-topped region due to the increasing effect of gravity. For simplicity, we only consider the two asymptotic regimes. At the transition, the isoviscous current is characterized by $R \sim 4$, $h_0 \sim 2$ and $t \sim 10$. In the following, we consider a modified Peclet number $Pe_m = Pe St_m^{-1}$ which integrates the effect of crystallization for clarity.

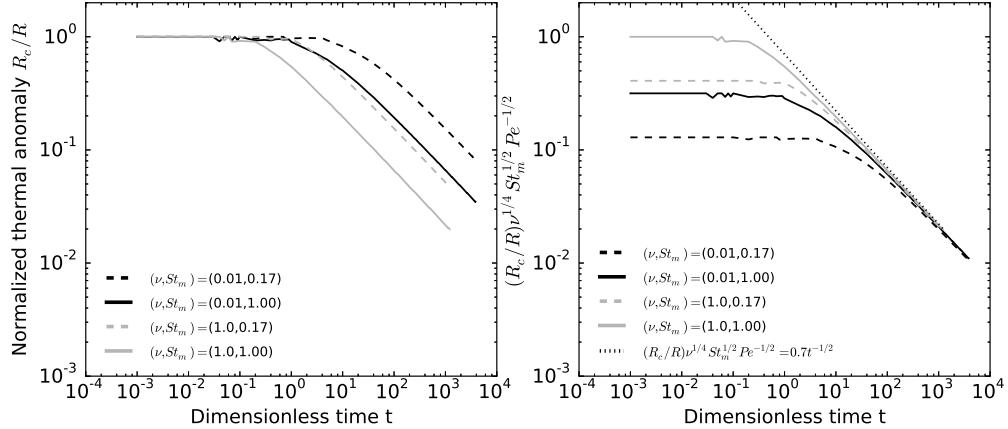


Figure 3.17: Left: Normalized thermal anomaly radius $R_c(t)/R(t)$ versus dimensionless time for different combinations (ν, St_m) indicated on the plot and $Pe = 1$. Right: Same plot but where we have rescaled the normalized thermal anomaly radius $R_c(t)/R(t)$ by $St_m^{-1/2} Pe^{1/2} \nu^{-1/4}$.

For a current with a temperature dependent viscosity, the transition between the bending regime and the gravity regime also occurs when the radius of the current is close to $R \sim 4$ (Figure 3.18). However, the time and thickness of the current at the transition depends on the thermal state of the flow, i.e. on the combination of (ν, Pe_m) considered (Figure 3.18). For instance, for $\nu = 0.01$ and a small value of Pe , i.e. $Pe = 1.0$ the current transitions to the gravity regime when it is in the third thermal phase of the bending regime, i.e. at $t \sim 50$ with $h_0 \sim 8$, as an isoviscous current with given cold viscosity $\eta_c = 100$. In contrast, for a larger value of Pe , i.e. $Pe = 10^5$, the current remains longer in the first phase of the bending regime and it spreads with hot viscosity η_h for a longer period of time. As a consequence, it reaches the transition sooner at $t \sim 30$ and with a smaller thickness $h_0 \sim 5$ when it is still in the second thermal phase of the bending regime. For an even larger Peclet number Pe , the current will transition in the first thermal phase of the bending regime at $t \sim 10$ and with $h_0 \sim 2$, as in the isoviscous case with viscosity η_h .

Overall, the time for the current to reach the transition t_t is the time for its radius to be larger than 4. It can be obtained from the scaling law followed by the radius $R(t)$ in the bending regime (3.83) and is equal to $6.5(\eta_e/\eta_h)^{2/7} h_f^{-1/7}$ where η_e is the effective viscosity of the current (see Section 3.4.5). In particular, it is bound by two values corresponding to two end-member cases: the case where the current transitions to the gravity regime while it is in the first bending phase, i.e. when $\eta_e = \eta_h$ and $t_h \sim 6.5h_f^{-1/7}$

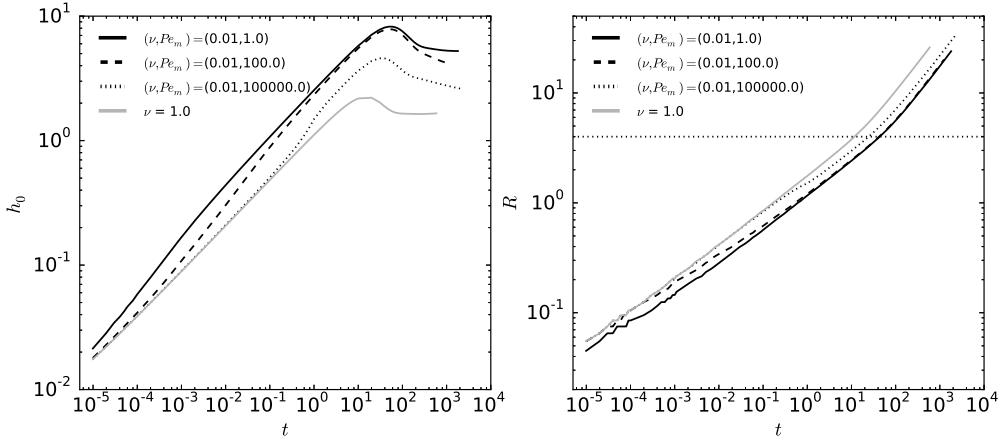


Figure 3.18: Left: Dimensionless thickness at the center h_0 versus dimensionless time for different sets (ν, Pe) indicated on the plot. The grey line represents the isoviscous case $\nu = 1$. Right: Same plot but for the dimensionless radius R . Horizontal black dotted-line represents the transition radius between the bending and the gravity regime.

and the case where the current transitions to the gravity regime while it is in the third bending phase, i.e. $\eta_e = \eta_c$ and $t_c \sim 6.5\nu^{-2/7}h_f^{-1/7}$. Indeed, when rescaling the time of the simulation by t_c , the different combinations (ν, Pe_m) , for which the third thermal phase of the bending regime has been reached before the transition to the gravity regime, collapse on the same curve (Figure 3.19, right).

The subsequent evolution in the gravity regime also depends on the combinations (ν, Pe_m) considered. Indeed, in contrast to the bending regime where the effective viscosity is that of a small region at the tip, the effective viscosity is the average flow viscosity in the gravity regime. Therefore, the effective viscosity of the flow can drastically decrease when entering the gravity regime and a flow in the i th thermal phase of the bending regime can transition in the j th thermal phase of the gravity regime with $i \geq j$ which results in 6 possible scenarios (see Appendix 3.8 for more details). For instance, a current in the second thermal phase of the bending regime can transition into the first or second thermal phase of the gravity current regime. However, the case where a current in the third thermal phase of the bending regime transitions to the first thermal phase of the gravity regime is not possible and in the following, we details the five remaining scenarios as a function of the combination (ν, Pe_m) considered.

We first consider the case where the current transitions to the gravity

Name	From	To	Expression
t_t	Bending	Gravity	$6.5(\eta_e/\eta_h)^{2/7}h_f^{-1/7}$
t_h	Bending	Gravity	$6.5h_f^{-1/7}$
t_c	Bending	Gravity	$6.5\nu^{-2/7}h_f^{-1/7}$
Bending regime			
t_{b2}	Phase 1	Phase 2	$0.1PeSt_m^{-1}h_f^2$
t_{b3}	Phase 2	Phase 3	$10^{-2}St_m^{-22/27}Pe^{22/27}\nu^{-14/27}h_f^{-7/27}$
Gravity regime			
t_{g2}	Phase 1	Phase 2	$10^{-2}PeSt_m^{-1}$
t_{g3}	Phase 2	Phase 3	$5PeSt_m^{-1}\nu^{-1/2}$

Table 3.1: Summary of the different transition times. t_t is the transition time between bending and gravity which is bound by t_h , when the current transitions in the first bending thermal phase, and t_c , when the current transitions in the third bending thermal phase. t_{b2} (resp. t_{b3}) represents the time to transition from phase 1 to phase 2 (resp. from phase 2 to phase 3) in the bending regime. t_{g2} (resp. t_{g3}) represents the time to transition from phase 1 to phase 2 (resp. from phase 2 to phase 3) in the gravity regime.

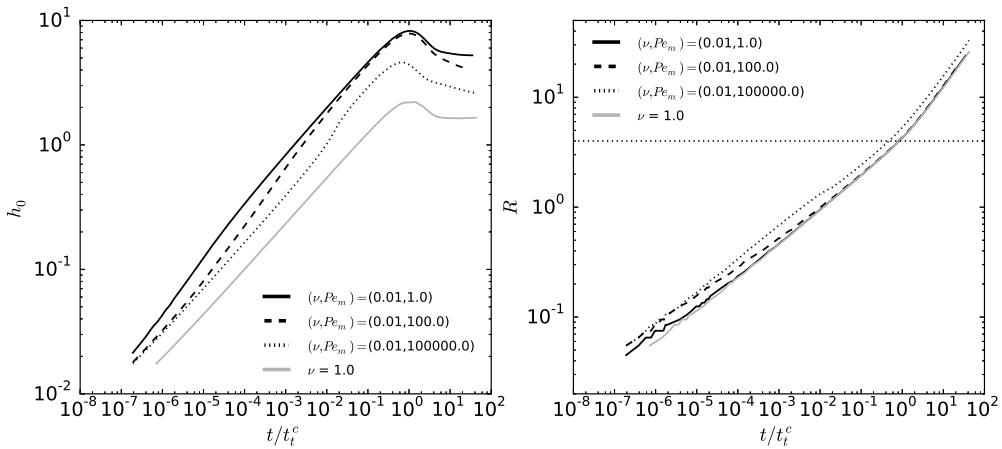


Figure 3.19: Left: Dimensionless thickness at the center h_0 versus time where the time has been rescaled by the time t_c the current transitions to the gravity regime while it is in the third bending phase (Table 3.1). The grey line represents the isoviscous case with given viscosity η_h . Right: Same plot but for the dimensionless radius R . Horizontal black dotted-line represents the transition radius between the bending and the gravity regime.

regime in the first thermal phase of the bending regime. In that case, the time for the transition is t_h and is less than the time for the second bending thermal phase change t_{b2} ; comparing t_h and t_{b2} gives $Pe > 65h_f^{-15/7}$ (Figure 3.20, Table 3.1). As $t_h < t_{g2}$ for $Pe > 65h_f^{-15/7}$, the current then transitions to the first thermal phase of the gravity current regime (B_1G_1 in Figure 3.20 and Figure 3.22, Table 3.2 in Appendix 3.8).

At the opposite, if the current has reached the third thermal bending phase, the transition occurs at t_c and is necessarily larger than t_{b3} ; comparing t_c and t_{b3} gives $\nu > 8.3 \cdot 10^{-13} Pe_m^{7/2} h_f^{-1/2}$ (Figure 3.20, Table 3.1). As $t_c > t_{g2}$ for $\nu > 8.3 \cdot 10^{-13} Pe_m^{7/2} h_f^{-1/2}$, the current can either transition to the second or third thermal phase of the bending regime (Figure 3.22, Table 3.2 in Appendix 3.8). If it transitions to the second phase of the gravity regime, then comparing t_c and t_{g3} gives $\nu < 0.3 Pe_m^{14/3} h_f^{2/3}$ (B_3G_2 on Figure 3.20) and if it transitions to the third phase of the gravity current, then $\nu > 0.3 Pe_m^{14/3} h_f^{2/3}$ (B_3G_3 on Figure 3.20).

In the case where the transition occurs when it is in the second bending phase, the time for the transition is not exactly known. However, it is bounded by t_h and t_c and we can therefore predict some evolution scenarios. Indeed, the transition time is necessarily smaller than t_c . Therefore, if $t_c < t_{g2}$, i.e. $\nu > 7.0 \cdot 10^9 Pe_m^{-7/2} h_f^{-1/2}$, the transition time is also smaller than t_{g2} and the current transitions to the first gravity thermal phase (B_2G_1 on Figure 3.20). Similarly, if $t_h > t_{g2}$, i.e. $Pe_m < 650h_f^{-1/7}$, then the transition time is larger than t_{g2} and the current transitions to the second gravity current phase (B_2G_2 on Figure 3.20).

3.7 Application to the arrest of terrestrial laccoliths

At shallow depth in the upper crust, roof lifting is the dominant process by which magma makes room for itself, which leads to the formation of laccoliths by bending of the overlying strata (*Johnson and Pollard*, 1973; *Pollard and Johnson*, 1973). The isoviscous elastic-plated gravity current model has been used to study their formation and show that their bell-shaped morphology is consistent with their arrest in the bending regime (*Michaut*, 2011; *Bunger and Cruden*, 2011). However, their radius is too small to be fractured controlled and their arrest might be better explained by their cooling (*Michaut*, 2011).

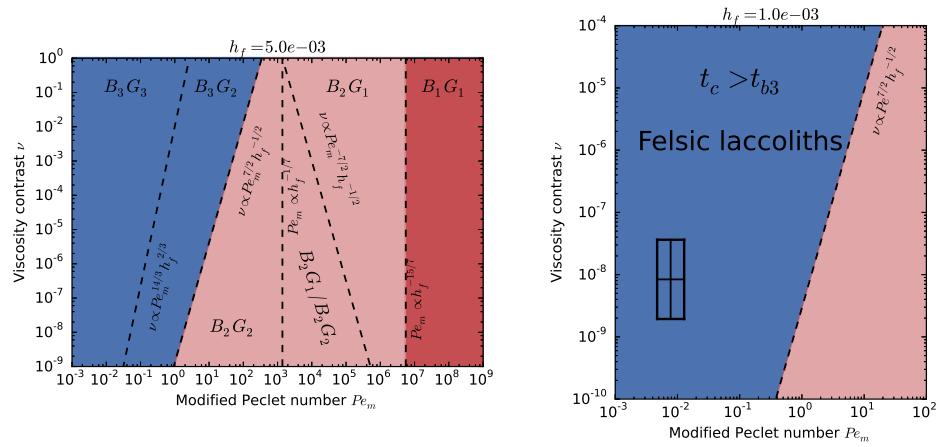


Figure 3.20: Left: Phase diagram for the evolution with bending and gravity for different combinations (ν, Pe_m) and a given value of $h_f = 0.005$. B_iG_j refers to the case where the current transitions from the i th bending thermal phase to the j th gravity thermal phase where i and $j \in \{1, 2, 3\}$. Right: Application to the spreading of laccoliths on a subset of the parameter space relevant for the study of terrestrial laccoliths described in Section 3.7.2. Rectangle: subset of the parameter space containing the laccoliths from *Rocchi et al.* (2002)

3.7.1 Range of values for the dimensionless numbers

For a lava density ρ_m of 2500 kg m^{-3} , Young's modulus values between 10 and 100 GPa and intrusion depths between 0.5 and 5 km, the characteristic length scale Λ varies between 1 and 10 km. The overpressure in magma reservoirs driving dykes is typically between a few to several tens of MPa (*Tait et al.*, 1988; *Martí*, 2000) and the conduit length Z_c varies from a few to several tens of kilometers. Lava viscosity at eruption temperature η_h depends mainly on its composition and water content; close to its liquidus temperature, it can vary from 10^3 to 10^6 Pa s for felsic lavas (*Shaw*, 1972; *Giordano et al.*, 2008; *Whittington et al.*, 2009; *Chevrel et al.*, 2013). Hence, injection rate Q_0 are in a range $0.01 - 1 \text{ m}^3 \text{ s}^{-1}$, the height scale H varies between 0.1 m and 2 m and the time scale τ varies from several months to a few years.

For a latent heat of crystallization $L = 4.18 \cdot 10^5 \text{ J kg}^{-1}$, a difference between solidus temperature T_S and liquidus temperature T_L between 100 K and 300 K, the number St_m varies from 0.1 to 0.5. For a thermal diffusivity for the magma equal to $\kappa_m = 10^{-6} \text{ m s}^{-2}$, the Peclet number can vary from 10^{-4} to 10 and therefore, Pem varies from 0.001 to 100. Finally, the increase in viscosity upon cooling can vary from 4 to 10 orders of magnitude (*Shaw*, 1972; *Lejeune and Richet*, 1995; *Giordano et al.*, 2008; *Diniega et al.*, 2013). We thus consider values of ν that vary between 10^{-4} and 10^{-10} .

The model also considers a thin pre-wetted film of thickness h_f whose meaning in the application to the spreading of laccolith is unclear. In particular, the model shows no convergence when h_f tends to zero (*Lister et al.*, 2013) and therefore, the thickness h_f might be linked to some structural length scale at the front of the laccolith or to the natural imperfection of the flow geometry. For the purpose of the application, we choose a film thickness of 1 mm, i.e. the minimum length scale with physical significance for the spreading of laccoliths which give a dimensionless h_f that varies between 10^{-2} and 10^{-4} . The limited effect of changing h_f is detailed in Appendix 3.8 and in the following, we set h_f to 10^{-3} .

It is generally assumed that magma stops spreading when its crystal content becomes close to its maximum packing, i.e. $\phi \sim 60\%$ (*Pinkerton and Stevenson*, 1992). Beyond this point, crystal collisions dominate and the viscosity jumps to much higher values (*Lejeune and Richet*, 1995; *Giordano et al.*, 2008). We assume that this is equivalent to η_e tending to η_c in our model. With this assumption, the model thus predicts that a magmatic intrusion would solidify as a laccolith upon reaching the third thermal phase of the bending regime. The phase diagram proposed in Section 3.6 simplifies and predicts that most felsic magmatic intrusions should indeed solidify as a laccolith (Figure 3.20, right).

3.7.2 Comparison with observations

Rocchi et al. (2002) provide data for the intrusion size and depth of nine laccoliths at Elba Island, Italy. The length and thickness scales can be estimated for each laccolith such that the data can be nondimensionalized and compare to the model. In particular, the thickness h_0 as a function of its radius R for a current that solidifies in the third phase of the bending regime can be derived from the scaling laws (3.82) and (3.83) and should follow

$$h_0 \sim 0.75\nu^{-2/7}R^{8/7} \quad (3.104)$$

For this example, each laccolith is part of a larger intrusive system, and hence variability of the model parameters should be limited, except for the overlying elastic layer thickness, taken to be the intrusion depth, whose variation between laccoliths is accounted for in the nondimensionalization. The observations show a very good agreement with the model for a viscosity contrast close to 8 orders of magnitude, which is consistent with the felsic composition of these laccoliths (Figure 3.21, left) (*Marsh*, 1981; *Diniega et al.*, 2013).

If the laccoliths have stopped spreading as soon as they reached the third phase of the bending regime, the variance in thickness and radius between the different intrusions should be explained only by variations in the Peclet number, most likely due to variations in intrusion depths in this example. Indeed, in Section (3.5.5), we show that the time t_{b3} the current reaches the third phase of the bending regime, and hence its thickness and its radius at this time, depends on the combination (ν, Pe_m) considered. To test this hypothesis, we estimate for each laccolith its Peclet number Pe using its corresponding depth of intrusion and use for the viscosity contrast $\nu = 8.2 \cdot 10^{-9}$ determined previously (Figure 3.20 and 3.21 right). Then, we rescale the variables using the time t_{b3} (3.93) as follow

$$\hat{t} = Pe_m^{-22/27}\nu^{14/27}t \quad \hat{R} = Pe_m^{-7/27}\nu^{2/27}R \quad \hat{h}_0 = Pe_m^{-8/27}\nu^{-10/27}h_0. \quad (3.105)$$

In term of \hat{h}_0 and \hat{R} , the scaling law (3.104) rewrites $\hat{h}_0 \sim 0.75\hat{R}^{8/7}$ and does not depend on the dimensionless numbers anymore. However, the different laccoliths do not collapse on the same dot after rescaling. In particular, the dependence of Pe of our scaling, resulting from different intrusion depths, is not enough to explain the variability in the size of terrestrial laccoliths. An additional cooling mechanism, depending on Pe , is thus required to explain the exact extent of laccoliths, which could be extraction of heat by circulation of fluid or heating of the wall rocks during the intrusion.

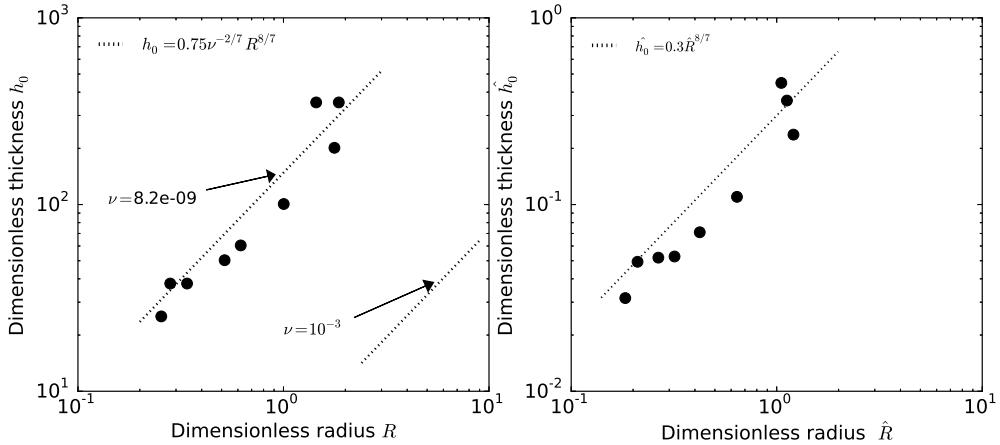


Figure 3.21: Left: Dimensionless maximum thickness h_0 versus radius R for laccoliths from Elba Island, Italy. Thickness, radius and depth for each laccolith are taken from [Rocchi et al. \(2002\)](#). Parameters for calculating Λ (3.22) and H (3.23) are $E = 10^9$ GPa, $\nu^* = 0.25$, $\rho_m = 2500$ kg m $^{-3}$, $g = 9.81$ m s $^{-2}$, $\eta_h = 10^6$ Pa s and $Q_0 = 0.1$ m 3 s $^{-1}$. For depth of intrusions between 1.9 and 3.7 km, the length scale Λ is between 2490 m and 3680 m. H is constant and equal to 1.98 m. Dotted lines: scaling laws (3.104) with $h_f = 0.001$ and two values for the viscosity contrast ν indicated on the plot. $\nu = 8.2 \cdot 10^{-9}$ represents the least square best fit for the data. Right: Dimensionless thickness \hat{h}_0 versus \hat{R} where \hat{h}_0 and \hat{R} are given by (3.105). Substituting (3.24) into (3.32), we obtain $Pe = Q_0H/(\pi\kappa\Lambda^2)$; the parameters for calculating Pe for each laccolith are the same than those used for the nondimensionalization, $\kappa = 10^{-6}$ m s $^{-2}$ and St_m is considered constant and set to 1. The viscosity contrast is set to $\nu = 8.2 \cdot 10^{-9}$ for all laccoliths. Dotted line: scaling law $\hat{h}_0 \sim 0.75\hat{R}^{8/7}$.

3.8 Summary and conclusion

Isothermal elastic-plated gravity current shows two asymptotic regimes. At early times, the gravity is negligible and the peeling of the front is driven by the bending of the overlying layer. In contrast, at late times, the own flow weight becomes the driving pressure and the current evolves in a so called gravity current regime. In this study, we have developed a theory for the evolution of an elastic-plated gravity current with a temperature dependent viscosity. In particular, we study the response of the flow to its cooling in each regime separately.

Scaling analyses of the heat transport equation show that the evolution of the thermal structure depends on the regime considered. In the bending regime, since the flow constantly thickens, the thermal anomaly grows with time but slower than the flow itself and a region of cold fluid rapidly forms at the front. The size of the cold fluid region depends on the dimensionless parameters of the system, i.e. the Peclet number, the viscosity contrast and a dimensionless number that accounts for crystallization. In contrast, in the gravity current regime, since the flow tends to a constant thickness, the temperature profile diffuses to an almost stationary profile and the thermal anomaly reaches a steady-state. The time for reaching this steady-state also scales with the dimensionless numbers of the system.

Numerical analyses of the equations show that the combine effect of cooling and temperature-dependent viscosity result in important deviations from the isoviscous case. In particular, each regime is split in three different phases. A first phase where the flow behaves as an isoviscous flow with hot viscosity. A second phase where the flow slows down and drastically thickens. A last phase where the flow returns in an isoviscous flow but with cold viscosity. These three phases are linked to the coupling between the thermal anomaly and the flow itself and in particular, the second phase of the flow is triggered by the detachment of the thermal anomaly. However, we show that the effective viscosity of the flow is drastically different in the two regimes. While the dynamics is governed by the local thermal condition at the front in the bending regime, it is the average thermal structure of the current that controls the flow in the gravity regime.

The final evolution of an elastic-plated gravity current therefore depends on the relative phase change within each regime and the transition between the bending and the gravity regime itself. We provide a general phase diagram that predicts the different evolution scenarios as a function of the dimensionless parameters. We finally apply the results to the spreading of magmatic intrusions and show that, if cooling is indeed an efficient mechanism for the arrest of laccolith in the bending regime, as confirmed by observations, an

additional cooling mechanism is needed to explain the exact size of laccoliths.

Appendix A: Numerical scheme

The coupled nonlinear partial differential equations (3.45) and (3.46) are solved on a grid much larger than the flow itself and shifted at the center to avoid problems arising from the axisymmetrical geometry. The procedure used to solve both equations, (3.45) and (3.46), is to use a finite-difference scheme for spatial discretization coupled with an implicit backward Euler scheme in time. In addition, since each equation is non-linear, we use Newton-Raphson method to iterate towards the solution at each time step for both equations. We begin the computation with $h = h_f$, $\Theta_b = 1$ and $\delta = 10^{-4}$ over the whole domain. In addition, we impose

$$\left. \frac{\partial h}{\partial r} \right|_{r=0} = \left. \frac{\partial P}{\partial r} \right|_{r=0} = 0 \quad (3.106)$$

and $h = h_f$ at the end of the grid.

The expressions of $I_0(\delta)$, $I_1(h)$, $I_1(\delta)$ and $I_2(\delta)$ are the following

$$I_0(\delta) = -\frac{\delta}{12} (-6\delta\nu + (1-\nu)(-5\Theta_b\delta + 4\Theta_bh) + 6h\nu) \quad (3.107)$$

$$I_1(h) = \frac{1}{60} ((1-\nu)(-4\Theta_b\delta^3 + 10\Theta_b\delta^2h - 10\Theta_b\delta h^2 + 5\Theta_bh^3) + 5h^3) \quad (3.108)$$

$$I_1(\delta) = -\frac{\delta^2}{120} (-40\delta\nu + (1-\nu)(-36\Theta_b\delta + 25\Theta_bh) + 30h\nu) \quad (3.109)$$

$$I_2(\delta) = -\frac{\Theta_b\delta^2}{2520} (-882\delta\nu + (1-\nu)(-778\Theta_b\delta + 560\Theta_bh) + 735h\nu) \quad (3.110)$$

and therefore, (3.47) and (3.48) reduce to

$$\bar{u} = -\frac{\delta}{10} (-20\delta\nu + (1-\nu)(-14\Theta_b\delta + 15\Theta_bh) + 30h\nu) \quad (3.111)$$

$$\Sigma = -\frac{\Theta_b\delta^2}{630} (-2058\delta\nu + (1-\nu)(-1784\Theta_b\delta + 1330\Theta_bh) + 1785h\nu) \quad (3.112)$$

Appendix B: Phase transitions

A current in the i th thermal phase can transition in the j th phase of the gravity regime. In the following, we show that a transition from i to j where $i < j$ is not possible. Indeed, the effective viscosity is that of a small region at the tip in the bending regime whereas it is the average viscosity of the flow in the gravity regime. Therefore, it cannot increase during the transition and

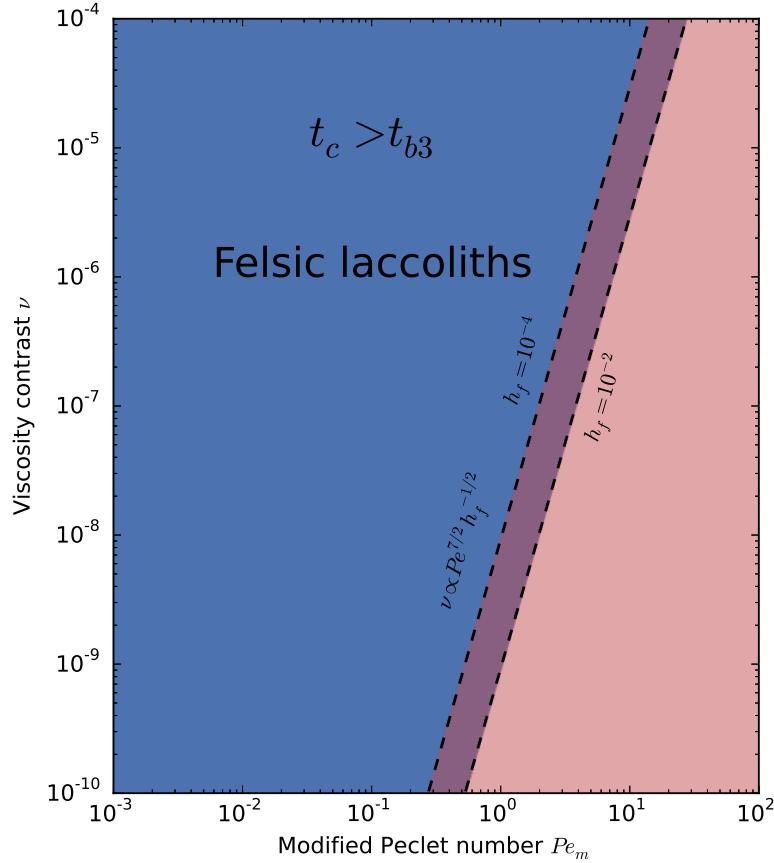


Figure 3.22: Phase transitions reported in Table 3.2

indeed, B_1G_2 , B_1G_3 and B_2G_3 are unfeasible (Table 3.2 and Figure 3.22). In addition, the transition from the third thermal phase of the bending regime to the first thermal phase of the gravity regime implies that $t_c < t_{b2}$ and $t_c > t_{g3}$, which is not possible (Table 3.2 and Figure 3.22). Therefore, the five possible sequences that remain are B_1G_1 , B_2G_1 , B_2G_2 , B_3G_2 and B_3G_3 (Table 3.2 and Figure 3.22).

Appendix C: Effect of the pre-wetted film thickness

The divergence of the viscous stresses at the contact line imposes the need for a regularization condition at the front (*Lister et al.*, 2013; *Flitton and King*, 2004; *Hewitt et al.*, 2014). In this study, we show the results of the simulations for a thin pre-wetted film of constant dimensionless thickness $h_f = 5 \cdot 10^{-3}$.

Transition	Condition 1	Condition 2	Condition 3	Output
Transition in the first bending thermal phase $B1$				
$t_t = t_h$	$t_h < t_{b2}$ $Pe_m > 65h_f^{15/7}$	$t_h < t_{g2}$ $Pe_m > 650h_f^{-1/7}$	-	B_1G_1
			-	Feasible
$t_t = t_h$	$t_h < t_{b2}$ $Pe_m > 65h_f^{15/7}$	$t_h > t_{g2}$ $Pe_m < 650h_f^{-1/7}$	$t_h < t_{g3}$ $\nu < 0.6Pe_m^2 h_f^{2/7}$	B_1G_2
				Unfeasible
$t_t = t_h$	$t_h < t_{b2}$ $Pe_m > 65h_f^{15/7}$	$t_h > t_{g3}$ $\nu > 0.6Pe_m^2 h_f^{2/7}$	-	B_1G_3
			-	Unfeasible
Transition in the second bending thermal phase $B2$				
$t_h < t_t < t_c$	$t_h > t_{b2}$ $Pe_m < 65h_f^{-15/7}$	$t_c < t_{b3}$ $\nu < \alpha Pe_m^{7/2} h_f^{-1/2}$	$t_c < t_{g2}$ $\nu > \beta Pe_m^{-7/2} h_f^{-1/2}$	B_2G_1
				Feasible
$t_h < t_t < t_c$	$t_h > t_{b2}$ $Pe_m < 65h_f^{-15/7}$	$t_c < t_{b3}$ $\nu < \alpha Pe_m^{7/2} h_f^{-1/2}$	$t_c < t_{g3}$ $\nu < 0.3Pe_m^{14/3} h_f^{2/3}$	B_2G_2 or B_2G_1
				Feasible
$t_h < t_t < t_c$	$t_h > t_{b2}$ $Pe_m < 65h_f^{-15/7}$	$t_c < t_{b3}$ $\nu < \alpha Pe_m^{7/2} h_f^{-1/2}$	$t_h > t_{g2}$ $Pe_m < 650h_f^{-1/7}$	B_2G_2
				Feasible
$t_h < t_t < t_c$	$t_h > t_{b2}$ $Pe_m < 65h_f^{-15/7}$	$t_c < t_{b3}$ $\nu < \alpha Pe_m^{7/2} h_f^{-1/2}$	$t_h > t_{g3}$ $\nu > 0.6Pe_m^2 h_f^{2/7}$	B_2G_3
				Unfeasible
Transition in the third bending thermal phase $B3$				
$t_t = t_c$	$t_c > t_{b3}$ $\nu > \alpha Pe_m^{7/2} h_f^{-1/2}$	$t_c < t_{g2}$ $\nu > \beta Pe_m^{-7/2} h_f^{-1/2}$	-	B_3G_1
			-	Unfeasible
$t_t = t_c$	$t_c < t_{b2}$ $\nu > \alpha Pe_m^{7/2} h_f^{-1/2}$	$t_c > t_{g2}$ $\nu < \beta Pe_m^{-7/2} h_f^{-1/2}$	$t_c < t_{g3}$ $\nu < 0.3Pe_m^{14/3} h_f^{2/3}$	B_3G_2
				Feasible
$t_t = t_c$	$t_c < t_{b2}$ $\nu > \alpha Pe_m^{7/2} h_f^{-1/2}$	$t_c > t_{g3}$ $\nu > 0.3Pe_m^{14/3} h_f^{2/3}$	-	B_3G_3
			-	Feasible

Table 3.2: Parameter space analysis. The coefficients α and β are $\alpha = 8.3 \cdot 10^{-13}$ and $\beta = 7.0 \cdot 10^9$

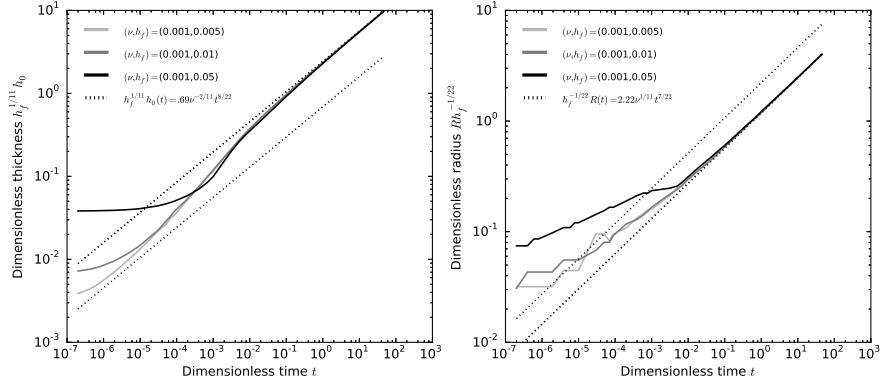


Figure 3.23: Left: Dimensionless thickness at the center $h_0 h_f^{2/22}$ versus dimensionless time t for different sets (ν, h_f) indicated on the plot. Dashed-lines represent the scaling laws $h_0 h_f^{2/22} = 0.7\nu^{-2/11} t^{8/22}$ for $\nu = 1.0$ and 0.001 . Right: Dimensionless radius R versus dimensionless time t for the same sets (ν, h_f) . Dashed-lines represent the scaling laws $R h_f^{-1/22} = 2.2\nu^{1/11} t^{7/22}$ for $\nu = 1.0$ and 0.001 .

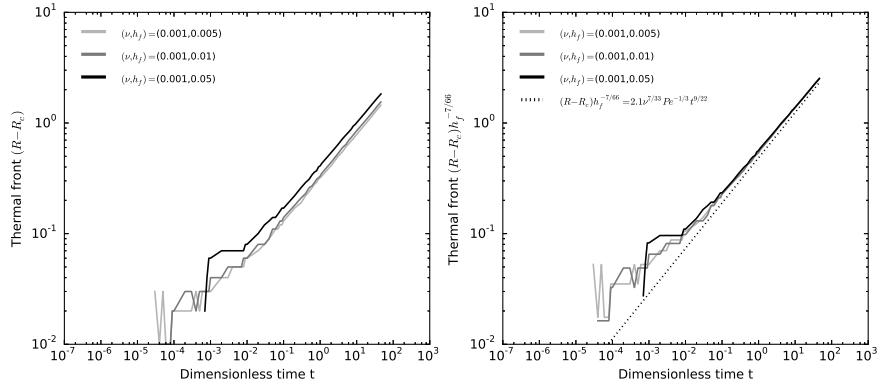


Figure 3.24: Left: Extent of the cold fluid region $R(t) - R_c(t)$ versus dimensionless time for different combinations (ν, h_f) indicated on the plot. Right: Same plot but where we have rescaled the extent of the cold fluid region by $h_f^{7/66}$. Dashed-line: scaling law $(R(t) - R_c(t)) h_f^{-7/66} = 2.1 P e^{-1/3} \nu^{7/33} t^{9/22}$.

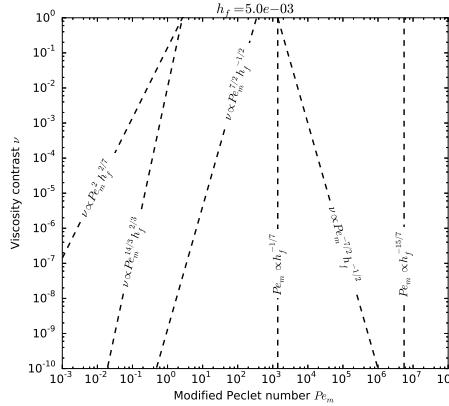


Figure 3.25: Phase diagrams for the evolution with bending and gravity for different combinations (ν, Pe_m) and different values for the film thickness $h_f = 10^{-2}, 10^{-3}$ and 10^{-4} .

However, the different scaling laws derived in Section 3.4 depends on the film thickness h_f , as confirmed numerically (Figure 3.23 and 3.24) and in particular, the phase diagram presented in section 3.6 and its application to the spreading of laccolith thus depends on the chosen value for h_f . The meaning of the pre-wetted film thickness in the application to the spreading of laccolith is unclear, however, reasonable values for h_f should range from a few centimeters to no less than 0.1 milimeter $\sim 10^{-4}$. Indeed the height scale has to be linked to some structural length scale at the front of the laccolith. Therefore, as the dependence with h_f is weak, a variation of 2 orders of magnitude does not change significantly the results (Figure 3.25).

The same result hold when we look at the relation between the thickness and the radius of the laccolith (3.104). Indeed, the best fit for the value of the viscosity contrast scales as $h_f^{-1/2}$, i.e. $\nu_{\text{best}} = h_f^{-1/2} 2.59 \cdot 10^{-10}$ and therefore, varying h_f by two orders of magnitudes change the viscosity contrast by one order of magnitude which is acceptable for our application.

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Part III

Cratères à sol fracturée: Témoins
d'un magmatisme intrusif lunaire

