Volatiles in volcanic systems

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Magmas contain only a small mass fraction of volatiles, yet its role in magma dynamics and eruption style is fundamental. Magmatic volatiles partition between liquid, solid and gas phases and in doing so change magma bulk density and compressibility, which have consequences for magma buoyancy and volume. An exsolved fluid phase, which may be distributed unevenly through reservoirs, contains sulfur and metals, which may be transported into the atmosphere or into porphyry deposits. We review the controls on volatile solubility and the methods to reconstruct the volatile budget of magmas, focussing on the exsolved gas phase. We consider the role of exsolved and evolving fluids on magma dynamics and on eruption style.

INTRODUCTION

Volcanic eruptions, in all their diverse forms, are driven by overpressure, buoyancy and degassing. Subaerial eruptions are associated with vast clouds of magmatic gases, a process that has shaped our hydrosphere over Earth's history. Volcanic eruption deposits, when viewed at high magnification using a scanning electron microscope, are dominated by vesicles: bubbles frozen in place by quenching during eruption (**figure 1**). There can be no doubt that in order to understand volcanic processes, the mechanisms and consequences of magmatic volatile degassing are paramount.

Magmatic volatiles are chemical constituents in melts that have relatively low molecular masses and partition into a gas phase at low pressures. The primary magmatic volatiles that drive volcanic eruptions are water (H_2O) and carbon dioxide (CO_2). Volatiles are only present in small amounts (a few weight per cent) in magmas, but their influence belies their weight. The presence of dissolved volatiles in silicate melts controls the presence, abundance and composition of crystal phases that grow in the magma during cooling and decompression; e.g. hydrous phases such as amphibole and plagioclase with more Ca-rich compositions crystallize in melts with high dissolved H_2O. In addition, the density of hydrous melts is lower relative to anhydrous melts

(Ochs III and Lange, 1999), which may promote magma ascent through the crust (owing to greater buoyancy) as well as convection in magma reservoirs.

Close to the Earth's surface, in the crust, melts typically reach saturation with respect to vapor (or gas) in silicate melts, such that volatiles exsolve to form a gas phase disseminated in the magma as bubbles. Exsolution may occur through "second boiling" in magma chambers – a result of prolonged cooling and crystallization – and this increases pressure in the chamber (because the surrounding country rocks are relatively incompressible) and may trigger eruptions (Tait et al., 1989). Volatile exsolution also results from the decompression that occurs during magma ascent to the surface ("first boiling"), and this exsolution causes tremendous expansion of bubbly magma, causing it to accelerate up the conduit. The rate and efficiency of exsolution and gas loss during ascent, as well as the rheological properties of the magma, largely control eruption style.

The vapor phase (referred to as "gas" at low pressures, and "fluid" at high pressure) in magma reservoirs in the crust is predominantly made up of CO_2 at depth, becoming more H_2O-rich at lower pressures. Magmatic gases also contain a myriad of other chemical species; chief of which are sulfur and halogens. Sulfur partitions strongly into the gas phase at low pressures (Scaillet et al., 1998). In relatively oxidised and lower temperature arc magmas in reservoirs in the upper crust, most of the sulfur in the magma is likely to exist in the gas phase, rather than dissolved in melt (Wallace, 2001). A consequence of this is that explosive volcanic eruptions, which evacuate these reservoirs rapidly, typically release far more sulfur (as sulfur dioxide) than can be accounted for by the amount dissolved in melt inclusions (tiny fragments of melt trapped in growing crystals at depth, prior to eruption). Chlorine, which also partitions into fluids, forms complexes with metals, a key part of the process of formation of Cu porphyry and Au epithermal deposits in the shallow plumbing systems of some arc volcanic systems (Williams-Jones and Heinrich, 2005).

The presence of vapor bubbles in magma changes its bulk physical and rheological properties in important ways. In particular, bubbles make magma compressible. Compressible magma responds to injections of new magma, or evacuation of magma during eruption, by contracting or expanding in response, effectively behaving like a "magma sponge" (Rivalta and Segall, 2008), and this behaviour has consequences for eruption longevity and duration (Huppert and Woods, 2002), and volcano monitoring (see Biggs and Pritchard, this issue). The exsolution of H_2O from melts removes a "network modifier", resulting in the lengthening of chains of silica polyhedra in the melt, increasing melt viscosity. The interplay between exsolution, viscosity change and outgassing (physical separation of the gas from the magma) during magma ascent is critical for determining eruption style (Dingwell et al., 1996).

The behavior of exsolved gases in long-lived, vertically-extensive, mush-dominated magmatic systems beneath volcanoes remains poorly understood and is the focus of much recent research. Gas bubbles may be retained in crystal-rich magmas by capillary forces at low gas fractions, may be transported through quasi-brittle fractures at high gas fractions, and might modify the bulk rheological properties of crystal mushes, rendering them able to respond (by mingling and reorganisation) on relatively fast timescales to magma recharge (Huber et al., 2011).

The reconstruction of volatile budgets in magmas that feed volcanic eruptions has received much attention from the perspective of understanding climate impacts (Thordarson and Self, 1993), mantle volatile systematics (Dixon and Clague, 2001), magma storage conditions (Wallace and Gerlach, 1994) and volcanic processes (Blundy and Cashman, 2005). We have various tools at our disposal for determining volatile concentrations in melts and their exsolution history, including melt inclusion geochemistry, the distribution of volatiles in solid phases, phase equilibrium experiments and thermodynamic models. Evaluating how exsolved fluids are generated and distributed in magma reservoirs, however, is more challenging. Owing to their buoyancy, gases may migrate and segregate from their source magmas. Long-lived reservoirs, which are subject to sporadic recharge and mingling of magmas, may develop complex reservoir architectures over time, with segregated regions of melts, mushes and exsolved gases (Christopher et al., 2015). In this article we review evidence for the formation, distribution and form of exsolved gases in magma reservoirs and their consequences for volcanic processes, both prior to and during eruptions.

THE ABUNDANCE AND DISTRIBUTION OF EXSOLVED GAS IN MAGMA RESERVOIRS

Vapor saturation of silicate melts occurs when the sum of the partial pressures of the dissolved volatiles in a melt is equal to the confining pressure, in which case a multicomponent gas phase will be in equilibrium with the magma. The solubility of the volatiles CO_2 and H_2O is mainly controlled by pressure (**figure 2**), and the much lower solubility of CO_2 compared to H_2O causes gases at higher pressures to be more CO_2 rich and those at lower pressures to be more H_2O rich. Bubbles are the consequence of vapor-saturation of magmas, and their nucleation and grow in silicate melts accommodates the exsolving vapor phase. Bubble nucleation may be homogeneous (in melt) or heterogeneous (on crystals) and requires volatile supersaturation to overcome surface tension. The extent of supersaturation is usually small and easily achieved in decompressing or crystallizing magmas, except in the case of homogeneous nucleation of bubbles in crystal-free rhyolitic magma, where strong melt supersaturation in volatiles may develop (Mangan and Sisson, 2000).

Other volatiles such as sulfur and halogens partition into the exsolved gas phase to varying extents. Sulfur and chlorine partitioning behaviour is well understood for a wide range of oxidation states and melt compositions (Zajacz et al., 2012). In general, experiments indicate that sulfur partitions strongly into the gas phase, particularly for more reducing conditions below the sulfate-sulfide transition (**figure 2**), caused by the lower solubility of sulfur when it exists as sulfide (S^{2-}) than when it occurs, under more oxidising conditions, as sulfate (S^{6+}). Saturation of the melt with Fe-sulfide melt or solid or with anhydrite (at more oxidising conditions) limits the sulfur concentration in the co-existing gas phase. Chlorine partitions much less strongly into a gas phase at magma chamber conditions than sulfur, and partitioning into the gas phase is much more pronounced for more silica-rich compositions (**figure 2**) (Zajacz et al., 2012). These chlorine-rich gases are important for transporting metals to the sites of hydrothermal ore deposits. Experimental data are consistent with the idea that for magmatic systems characterised by long-lived, evolved magmas recharged by underplating mafic magmas, the gas phase coexisting with the more evolved magma will be more chlorine-rich, whilst the gas phase supplied by the mafic magmas will be sulfur-rich, to a degree that depends on oxidation state and whether saturation with respect to a solid sulfur-bearing phase has been reached.

Observations constraining vapor-saturation of melts in the crust

Mafic, primitive melts in arc settings contain an average of 4 wt% H_2O (Plank et al., 2013) in melt inclusions and have been inferred to contain 3000 ppm to 1 wt% CO_2 (from melt inclusions and from modelling) (Blundy et al., 2010; Wallace, 2005) before significant differentiation occurs. Hotspot primitive basalts contain 0.5 to 1.5 wt% H_2O, and mid-ocean ridge basalts 0.3-0.5 wt% H_2O, based on analyses of submarine pillow-rim glasses and in melt inclusions (Dixon et al., 2002). The CO_2 contents of these primitive melts are less well constrained (due to degassing of melts prior to eruption and prior to entrapment as melt inclusions) but may reach levels similar to arc basalts, implying that vapor saturation may occur in some settings soon after melts are generated in the mantle and certainly by the mid-crust. The deep fluids released by magma intrusion in arcs have long been linked to flux melting and assimilation of lower crustal rocks and the generation of intermediate magmas (Annen et al.,

2006) as well as to the linked processes of lower crustal dehydration and formation of granulite terrains.

There is strong evidence that magmas stored in the mid to upper crust prior to eruption are usually vapor-saturated as a consequence of magma differentiation, prolonged storage, and recharge or underplating by CO_2-rich mafic magma (Wallace, 2001). For some volcanic systems, the geochemistry of melt inclusions has been used to reconstruct the exsolved vapor phase; chief among these is the Long Valley Caldera system in California, USA, through studies of melt inclusions hosted by quartz in the Bishop Tuff (Wallace et al., 1999). Volatiles and trace element concentrations in the melt inclusions are consistent with gas-saturated crystallization. Calculations of the amount of crystallization that took place, based on trace element variations in the melt inclusions, suggest that the magma chamber was zoned with respect to exsolved gas, varying from ~ 1 wt% exsolved gas near the bottom of the body (at ~ 250 MPa) to ~ 6 wt% near the top (at ~ 150 MPa) (Wallace et al., 1999).

Another approach to quantify the abundance of exsolved fluids in volcanic magma reservoirs is to compare the mass flux of volcanic gases with the flux of magma erupted. At Soufriere Hills Volcano, Montserrat, measurements of the gas composition and flux during the eruption (1995 to 2011) permitted estimates to be made of 2-8 wt% exsolved gas in the magma prior to eruption (Edmonds et al., 2014), similar to estimates from the Long Valley system (Wallace et al., 1999). This quantity of exsolved volatiles in the magma would impart a significant compressibility to the magma, which would result in only very muted volume changes in the magma reservior in response to either recharge or eruption (**figure 3**). This muted deformation is in fact observed: the volume decrease (deflation) during periods of eruption measured using a network of GPS receivers around the volcano is only around one tenth of the volume erupted (Elsworth et al., 2008). The presence of significant exsolved volatiles also has the effect of greatly increasing eruption longevity owing to the greater compressibility, leading to eruption of a larger mass to relieve the same overpressure (Huppert and Woods, 2002) (**figure 3**).

Observations of large, sulfur dioxide clouds accompanying explosive volcanic eruptions have been proposed to require a pre-eruptive gas phase (containing sulfur) in the magma reservoir (Wallace, 2001). Notably the eruptions associated with the largest sulfur clouds per erupted unit volume of magma appear to be intermediate arc magmas (andesites and dacites) (Wallace, 2001), such as the eruption cloud shown in **figure 3**, from Kasatochi Volcano, Alaska, in 2008. The degree to which sulfur partitions into the gas phase is maximised when the magmas are silica and H_2O-rich, relatively Fe and alkali-poor, and at low temperature, conditions met by the vast majority of explosive arc eruptions over the last few decades. The opportunity to observe a contrast to this sulfur-rich case may have presented itself recently: the plinian eruption of Chaiten Volcano, Chile, in 2008 was associated with a remarkably sulfur-poor cloud (Carn et al., 2009) accompanying the eruption of crystal-poor rhyolite. This low abundance of gaseous sulfur may have been due to the magma being relatively oxidised (i.e. sulfate-dominated), such that sulfur fluid-melt partitioning was inhibited in the presence of anhydrite, as suggested by recent modelling (Masotta et al., 2016). The presence of exsolved gas and the conditions under which it is generated therefore affects directly the sulfur loading of the atmosphere accompanying explosive eruptions. This is of great interest because sulfate aerosol interacts with solar radiation in the stratosphere, causing tropospheric cooling which, for the largest eruptions, may be severe, with mean global cooling of perhaps a few degrees Centigrade over a few years for VEI 6-7 eruptions, resulting in implications for climate and our environment.

Dynamics of gas-rich magma in the crust and eruption triggering

The fluid mechanics of bubbly magma in the crust and the role of bubbles in magma mixing, eruption triggering and the dynamics of magma reservoirs have been studied using analogue materials and numerical modelling. It has long been recognised that mafic magmas underplating more silicic, viscous magmas can produce a range of dynamical behaviors. If the mafic magma is vapor-saturated, bubbles may accumulate at the magma interface, lowering the bulk density of the mafic magma, which can induce overturn, mafic enclave formation, or bubble rise up through the interface, depending on the viscosity contrast between the two magmas. It has been suggested that the heating and remobilization of crystal-rich magma bodies in the arc crust may take place by "gas sparging", a process of mafic underplating, quenching and outgassing of a fluid phase that advects heat up through the pore spaces of a sub-solidus crystal-rich magma, thus partially melting it and perhaps triggering it to erupt on the timescales of a few months for relatively small-scale systems (10s of km^3 in size) (Bachmann and Bergantz, 2006).

A great deal of attention has been devoted to the microphysics of multi-phase crystal mushes in the crust and in particular how a gas phase might be trapped and/or might migrate through a crystal-rich mush. The presence of an exsolved gas phase has implications for bulk mush rheological properties: experiments have shown that the presence of only a few weight per cent of exsolved gas reduces the effective bulk viscosity of a crystal-rich magma substantially and induces shear-thinning behaviour, where the viscosity decreases with increasing strain rate. The presence of significant proportions of gas bubbles could make mushes more mobile in response to magma recharge events, potentially allowing large volumes of magma to be tapped in large eruptions; while the removal of such a gas phase (through outgassing) could result in "viscous death" and the formation of plutonic bodies. Experiments to investigate the mobility of such a gas phase through a partially molten mush show that bubbles that are of a similar size to the crystals may move upward through a mush under compaction by splitting and moving around crystals, whilst larger bubbles become trapped in the pore spaces. Injection of gases into crystal-rich suspensions in analogue materials shows that gas migration occurs by a range of processes, including viscous fingering and quasi-brittle failure, suggesting that under conditions of high gas or high crystal contents, mushes may behave in a brittle way with a yield strength, allowing gases to migrate relatively rapidly through them (**figure 4**) (Oppenheimer et al., 2015). These experiments show that crystal mushes and complex magmatic systems in the crust have the ability both to "hold" gas in the liquid layers but also to allow rapid outgassing under different conditions.

EXSOLUTION OF VOLATILES DURING ERUPTION

Eruptions may be triggered when overpressures exceed the tensile strength of the country rocks, allowing magma to ascend along fractures towards the surface. These overpressures may be caused by magma recharge or second boiling. As magma ascends, pre-existing bubbles will grow, or a new population of bubbles may nucleate (for high decompression rates), owing to the continued lowering of H_2O solubility in silicate melts at low pressures (**figure 2**). Bubble growth during magma decompression is limited by two main factors: the rate of diffusion of volatiles through melt into bubbles and the rate of viscous deformation of melt as bubbles expand. For high melt viscosities ($\sim>10^9$ Pas), viscous retardation may limit expansion of bubbles, causing the development of overpressure and perhaps Vulcanian explosive activity.

This growth of the gas phase driven by H_2O exsolution has multiple immediate consequences: the exsolution of H_2O from the silicate melt raises both the melt viscosity (Dingwell et al., 1996) and the solidus temperature, inducing a rapid burst of crystallization; and the magma's bulk density is drastically lowered as a result of gas bubble expansion, particularly in the uppermost few km of the conduit where pressure drops over several orders of magnitude, causing acceleration of the magma up the conduit (through conservation of mass). This combination of processes makes for a rich variety of possible volcanic eruption styles (**figure 5**), which are dependent on the interplay between magma decompression rate and the rheological properties of the melt. Interestingly, and somewhat counter-intuitively, the outcomes are not very sensitive to the total volatile content of the magma, except perhaps in the case of Strombolian activity (see below).

Influence of volatile exsolution and outgassing on volcanic eruption styles

For basaltic eruptions, which involve relatively low viscosity melt, eruption style is governed by ascent rate and the dynamics of two phase flow (Houghton et al., 2016). At low ascent rates, bubbles rise through melts in the conduit, inducing an eruptive spectrum from quiescent degassing to effusive activity to mild Strombolian eruptions, accompanied by persistent tropospheric gas plumes. This range of activity also broadly includes magma convection in a conduit, supplying volatiles to the atmosphere through outgassing of rising and bursting bubbles on the surface of the lava lake, followed by sinking of the denser degassed magma (Kazahaya et al., 1994).

Explosive styles of basaltic volcanic eruptions fall mainly into the categories of Hawaiian and Strombolian (**figure 5**). Observations, analogue experiments and textural studies (Houghton et al., 2016) demonstrate that these styles display a continuum in eruption intensity and magnitude, with Strombolian eruptions discrete in duration (typically < 100 seconds), with mass fluxes of 10^2-10^4 kg/s and Hawaiian eruptions more long-lived (typically > 2 hours), with mass fluxes of 10^4-10^6 kg/s (Houghton et al., 2016). The differences in duration and vigor are caused by fluctuations in eruption rate driven by the pressure regime in the magma chamber, which has non-linear consequences for two phase flow in the conduit. At low magma ascent rates, rising gas slugs and bubbles dominate, with their mass growth limited by diffusion; Strombolian activity may be related to the bursting of segregated single slugs or trains of bubbles. At Stromboli, Italy, it has been shown that more crystalline magma at the top of the conduit may act as a plug, promoting gas accumulation at shallow depths. Mixing between hotter, primitive melts and more evolved crystal-rich magma might be facilitated by the turbulent rise of gas bubbles and slugs (Lautze and Houghton, 2007). At higher magma ascent rates, continuous vesiculation and expansion causes inertia-dominated magma fragmentation (Namiki and Manga, 2008) and Hawaiian fountaining, which may persist for as long as overpressures in the reservoir remain elevated. Violent Strombolian activity, which is common at basaltic cinder cones, involves simulaneous explosive activity at the top of the cone and lava effuision from vents near the base. This style of activity occurs occurs in more H_2O-rich basaltic magmas at mass fluxes of 10^4-10^5 kg/s and is thus intermediate between Strombolian and sub-Plinian regimes (Pioli et al., 2008).

For high viscosity magmas, magma decompression rate (governed by magma chamber overpressure) and the rheological properties of the magma control the style of eruption. Here the viscous retardation of bubble growth generates overpressure in bubbles in silica-rich magmas. Magma fragmentation is driven by overpressure overcoming the tensile strength of the surrounding melt, with the fragmentation wave stopping at some critical overpressure threshold at depth in the conduit. For the case of Vulcanian eruptions, the high bulk viscosity of the magma precludes conduit refilling on the timescale of the eruption, rendering the eruption discrete in duration and limited in magnitude. For lower bulk viscosity magmas and large magma chamber overpressures, plinian eruptions, with magma column heights of 10s of km, often penetrating the stratosphere, are driven by continuous magma fragementation and magma flow, refilling the conduit on timescales of eruption (**figure 5**). The primary mode of magma fragmentation here might be brittle failure caused by rapid strain rates experienced by the rapidly vesiculating and expanding magma (Papale, 1999).

At low magma ascent rates, when melt relaxation can keep pace with bubble growth, permeable bubble networks can develop as the melt vesiculates. This gives magma the ability to effectively outgas volatiles both upward into the atmosphere and laterally into shear-fragmented conduit margins and country rocks, thus preventing magma fragmentation and explosive eruptions. It is thought that effective magma permeabilities for outgassing can develop at porosities of around 30% but perhaps at much lower porosities for sheared magmas (Rust and Cashman, 2004). Under these conditions, magmas erupt effusively in the form of steep-sided lava flows or domes.

FUTURE PERSPECTIVES

It has been proposed that andesitic volcanoes are essentially “ventholes” that allow excess subducted volatiles to be recycled to the surface, where the term “venthole” implies the presence and rise of a free volatile phase all the way down to the zones of arc magma generation (Giggenbach, 1996). While this may be an extreme view, the data for sulfur dioxide emissions from volcanoes discussed above suggests that in most cases, magma bodies in the mid to upper crust not only are vapor saturated but actually need to become gas charged before they can erupt, unless an eruption is “prematurely” triggered by tectonic or fault activity. Indeed, vapor saturation in crustal magma bodies can be seen as the inevitable consequence of the high CO_2 contents of most mantle-derived mafic magmas combined with the relatively low solubility of CO_2 at crustal pressures such that recharge or underplating by basaltic magma causes fluxing of CO_2-rich gas to occur (Wallace, 2003). The results of this can be seen at many scales – where the differences between systems reflect the distribution, geometry and size of magma bodies in the crust. Within this framework, long term periods of eruption and unrest at volcanoes such as Montserrat or Popocatépetl can be viewed as essentially intrusive events, in which the mass of erupted magma, while potentially devasting to the environment around the volcano, is quite small in comparison to the likely masses of both stored, differentiated magma and mafic, recently intruded magma at depth. Explosive eruptions of much larger magma bodies, by contrast, contain within them the exsolved gases released from underlying mush zones and recharging mafic magmas (Parmigiani et al., 2016), and might reflect thousands or even tens of thousands of years of gas accumulation (Christopher et al., 2015).

In light of this, the future for understanding volatiles in magmatic systems is to develop methods for tracking the movement of magmatic gas and fluids independent of magma movement. Methods for doing this include (1) tracking of the concentrations and fluxes of volatile components of very different solubilities, using melt inclusions and volcanic emissions, (2) phase equilibrium experiments and textural studies that can reveal processes such as gas fluxing, (3) seismic, ground deformation and other geophysical methods such as magnetotelluric data, and (4) use of volatile isotopes like ^{210}Po that can reveal information about the volumes and timescales of degassing magma bodies.

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Figures

Figure 1: **Photomicrographs of volcanic rocks illustrating the prevalence and importance of the exsolved gas phase in driving magma expansion and ultimately, volcanic eruptions.** The images show products from a diverse range of eruption types for which the style and extent of gas loss are a critical control on eruption style: a) backscattered electron image of pantelleritic pumice from Pantelleria, Italy (vesicles are black); b) transmitted light photomicrograph of a lava dome rock from Soufriere Hills Volcano, Montserrat, showing rounded white vesicles, tabular plagioclase, dark brown pleiochroic hornblende and light brown glass; c) backscattered electron image of scoria erupted at Stromboli, Italy. Large, coalesced vesicles are black, plagioclase phenocrysts grey and olivine (on left of image), white.

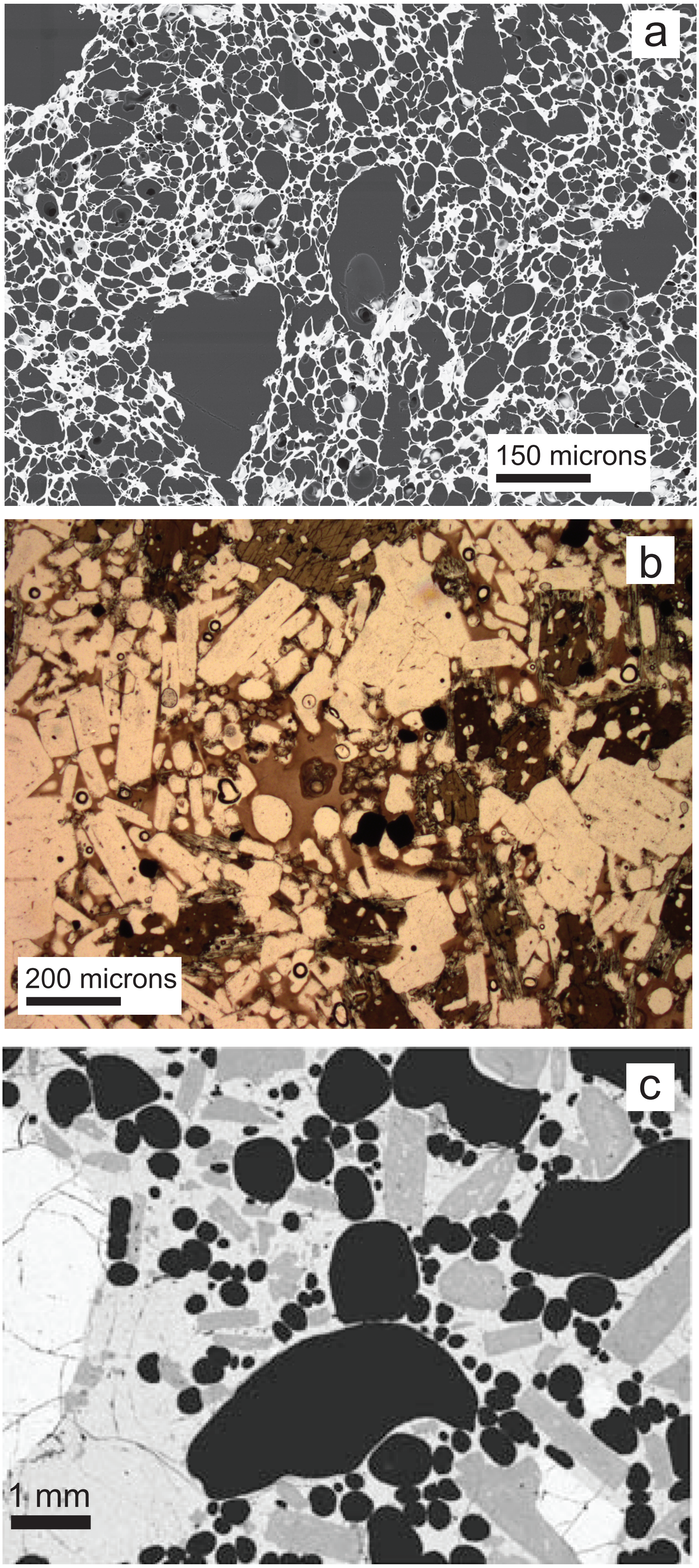
Figure 2: **Solubilities of volatiles in silicate melts and the controls on the development of an exsolved vapor phase**. a) H_2O-CO_2 saturation model, showing the CO_2 and H_2O concentrations stable in a silicate melt for a range of pressures (Newman and Lowenstern, 2002); b) The solubility of sulfur-bearing phases (in the absence of degassing) in silicate melt as a function of oxygen fugacity, showing the fields of typical MORB (in which dominantly sulfide is stable) and arc basalts (in which anhydrite becomes stable) (Jugo, 2009); c) schematic diagram illustrating the controls on the partitioning of sulfur between vapor and melt in silicate magmas, illustrating the importance of melt composition and oxidation state on the fluid-melt partition coefficient for sulfur; d) evolution of the melt volatile contents of a typical water-poor ocean island basalt and a water-rich arc dacite with pressure, showing the contrasing behaviour of sulfur in each case. The yellow shaded area is the typical depth range for the magma reservoirs feeding eruptions. Panel on right shows how the gas volume fraction evolves with decompression for each case.

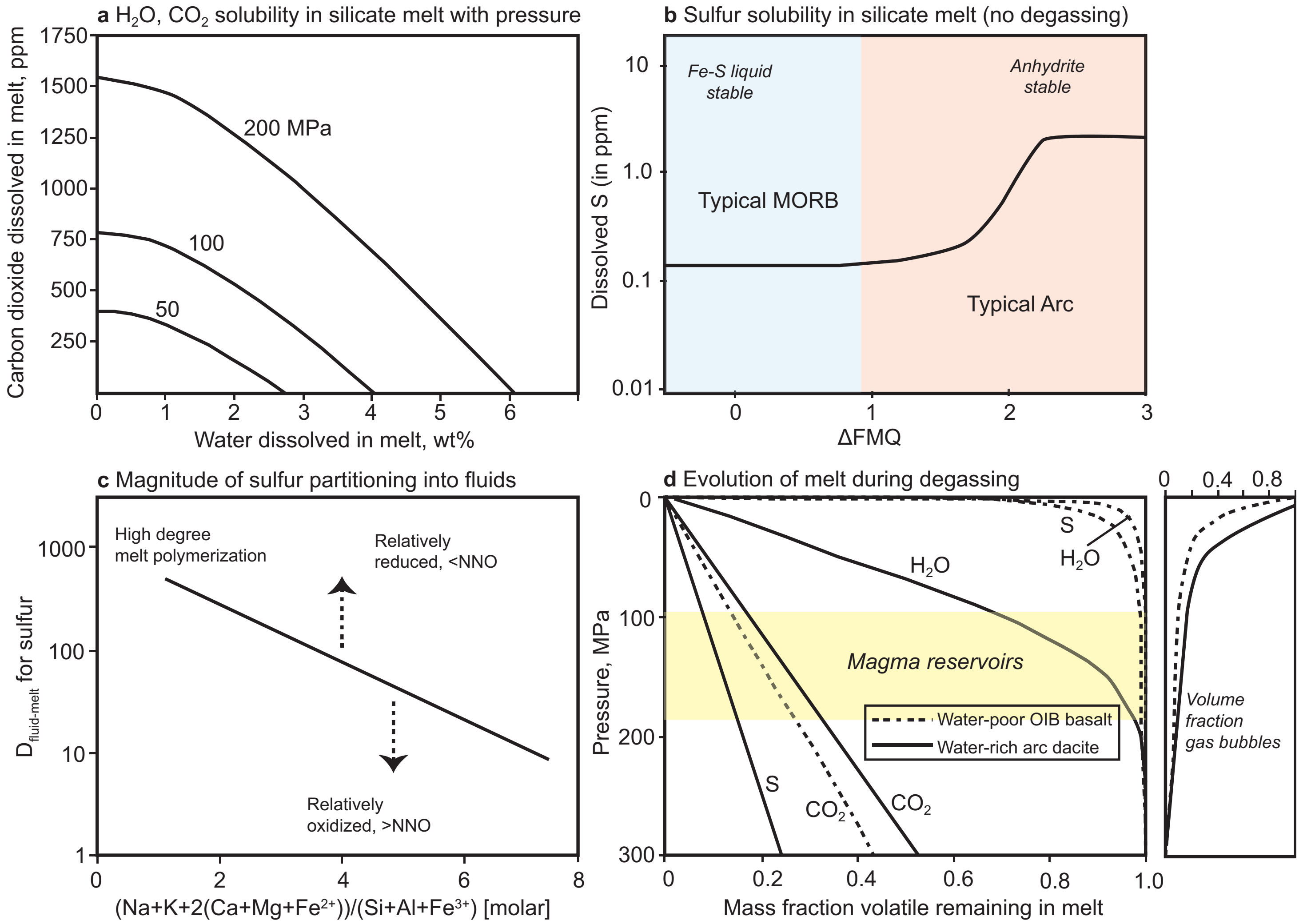
Figure 3: **Some physical and chemical consequences of gas-rich magma reservoirs**. a) volcanic eruptions are typically associated with large clouds of SO_2-rich gases. An AIRS image here shows the SO_2 cloud associated with the 2008 eruption of Kasatochi, Alaska (credit F. Prata); b) the presence of gas bubbles in magmas prior to eruption causes magma to be compressible, which results in only muted deformation being observed at the surface prior to and after eruptions, owing to the buffering effect of the compressible gas phase on volume in response to pressure.

Figure 4: Gas transport and storage in crystal-rich, recharging magma reservoirs, after (Parmigiani et al., 2016). a) a schematic of a magma reservoir, illustrating that gas may be

transported at a higher rate (see b) through crystal-rich mushy layers (by fracturing) than through liquid-rich layers (by buoyant bubble rise), leading to gas accumulation in liquid-rich regions. Eruptions tapping these liquid- and gas-rich regions would be associated with large gas clouds.

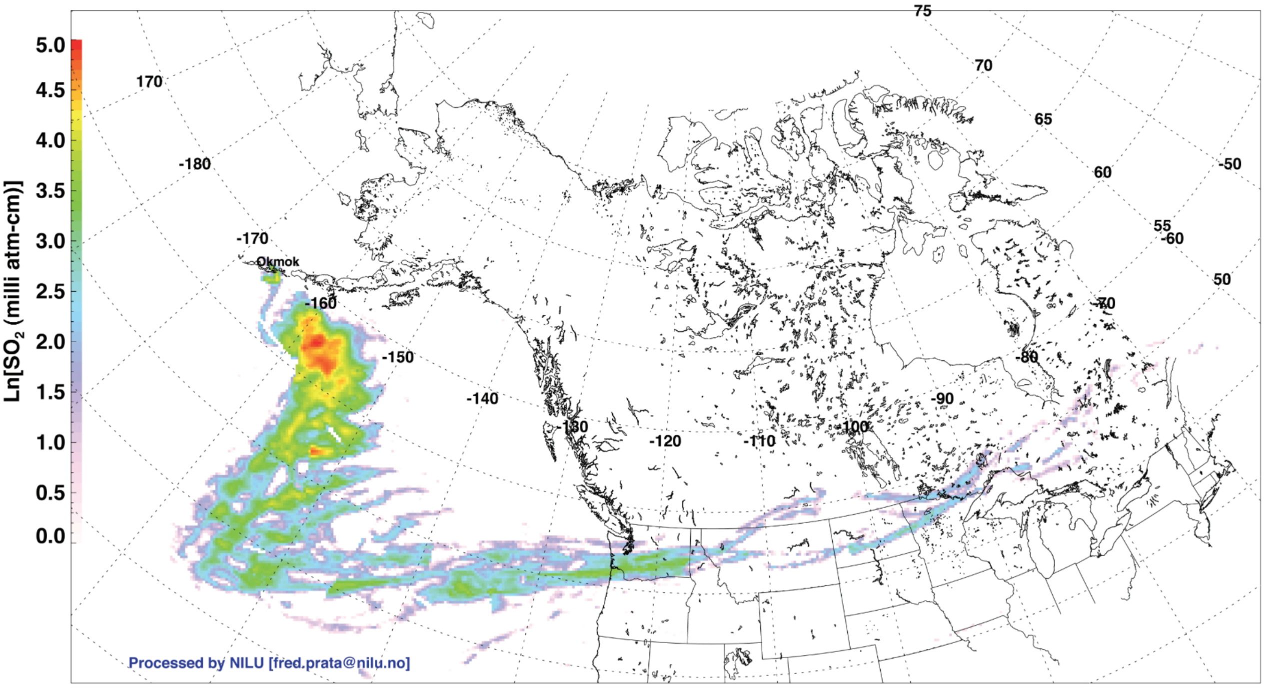
Figure 5: The effects of outgassing style and magma ascent rate on volcanic eruption style. Images show a representative range of volcanic eruption styles, with styles associated with rapid magma decompression at the top and slow magma decompression at the bottom. Viscosity increases from left to right.


a
150 microns
b
200 microns
c
1 mm

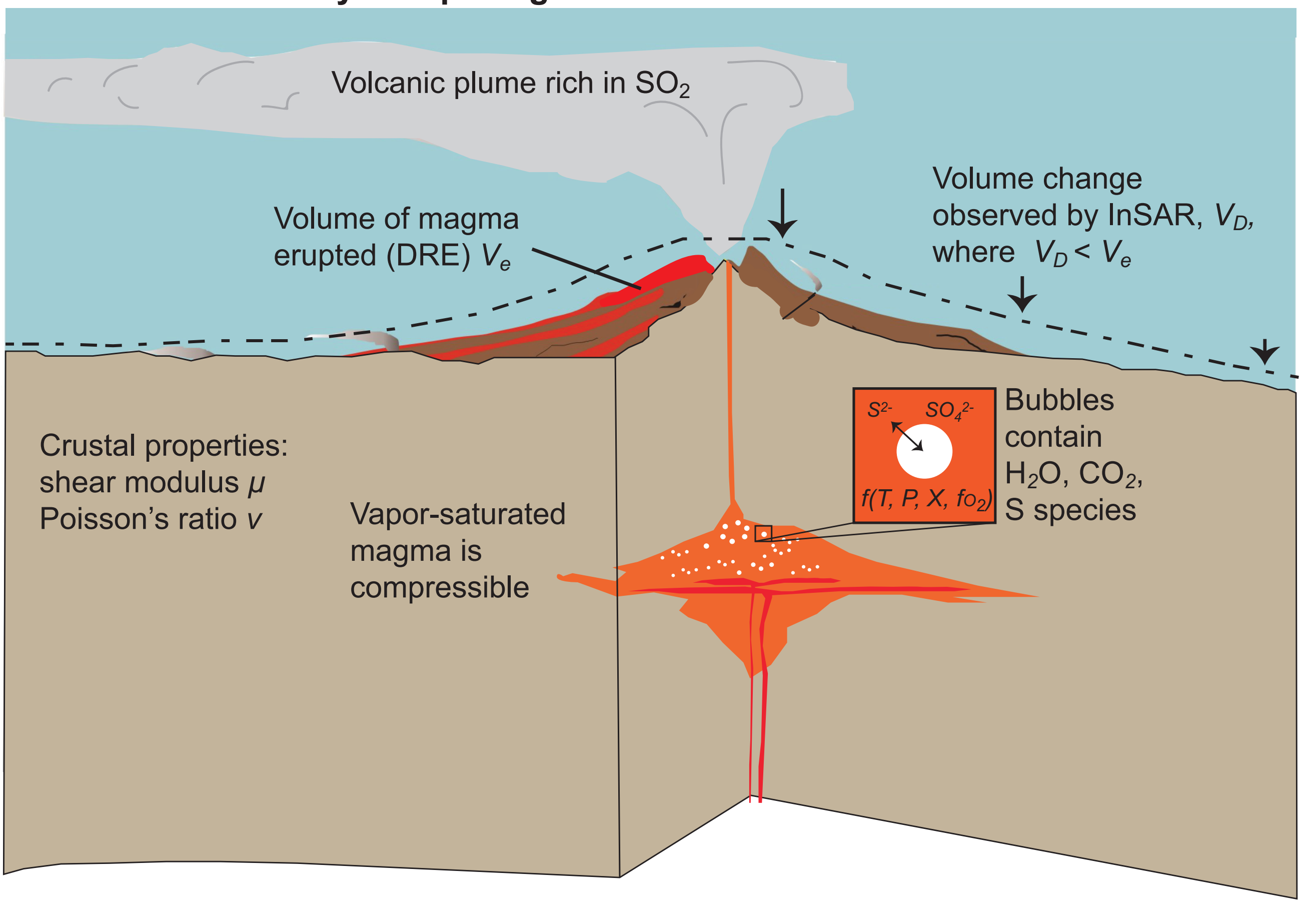


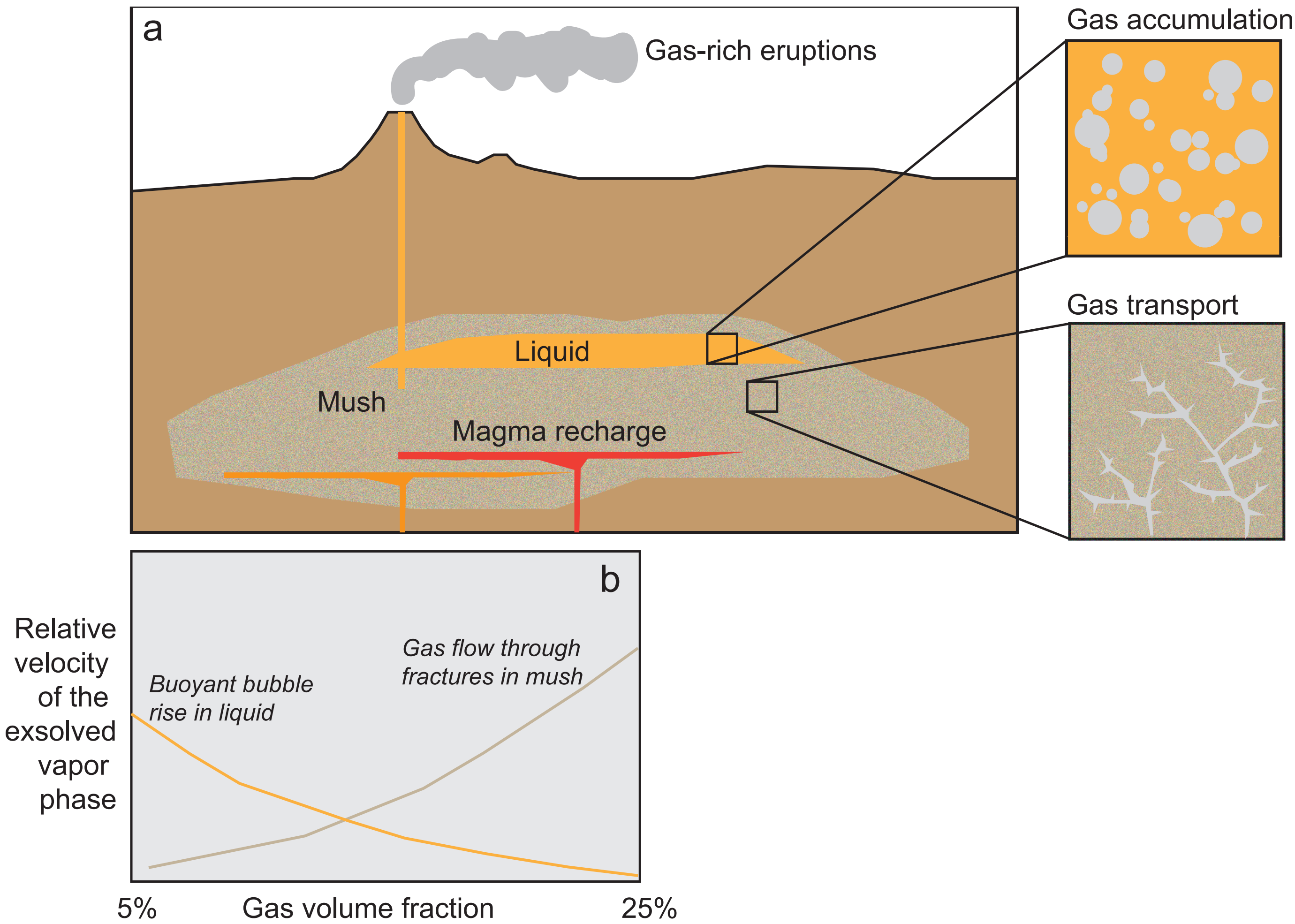
a H_2O, CO_2 solubility in silicate melt with pressure
Carbon dioxide dissolved in melt, ppm
1750
1500
1250
1000
750
500
250
200 MPa
100
50
0
1
2
3
4
5
6
Water dissolved in melt, wt%
b Sulfur solubility in silicate melt (no degassing)
Fe-S liquid stable
Anhydrite stable
Typical MORB
Typical Arc
Dissolved S (in ppm)
10
1.0
0.1
0.01
0
1
2
3
ΔFMQ
c Magnitude of sulfur partitioning into fluids
High degree melt polymerization
Relatively reduced, <NNO
Relatively oxidized, >NNO
$D_{fluid-melt}$ for sulfur
1000
100
10
1
0
2
4
6
8
(Na+K+2(Ca+Mg+Fe^{2+}))/(Si+Al+Fe^{3+}) [molar]
d Evolution of melt during degassing
Pressure, MPa
0
100
200
300
S
H_2O
H_2O
Magma reservoirs
Water-poor OIB basalt
Water-rich arc dacite
S
CO_2
CO_2
0
0.2
0.4
0.6
0.8
1.0
Mass fraction volatile remaining in melt
0
0.4
0.8
Volume fraction gas bubbles

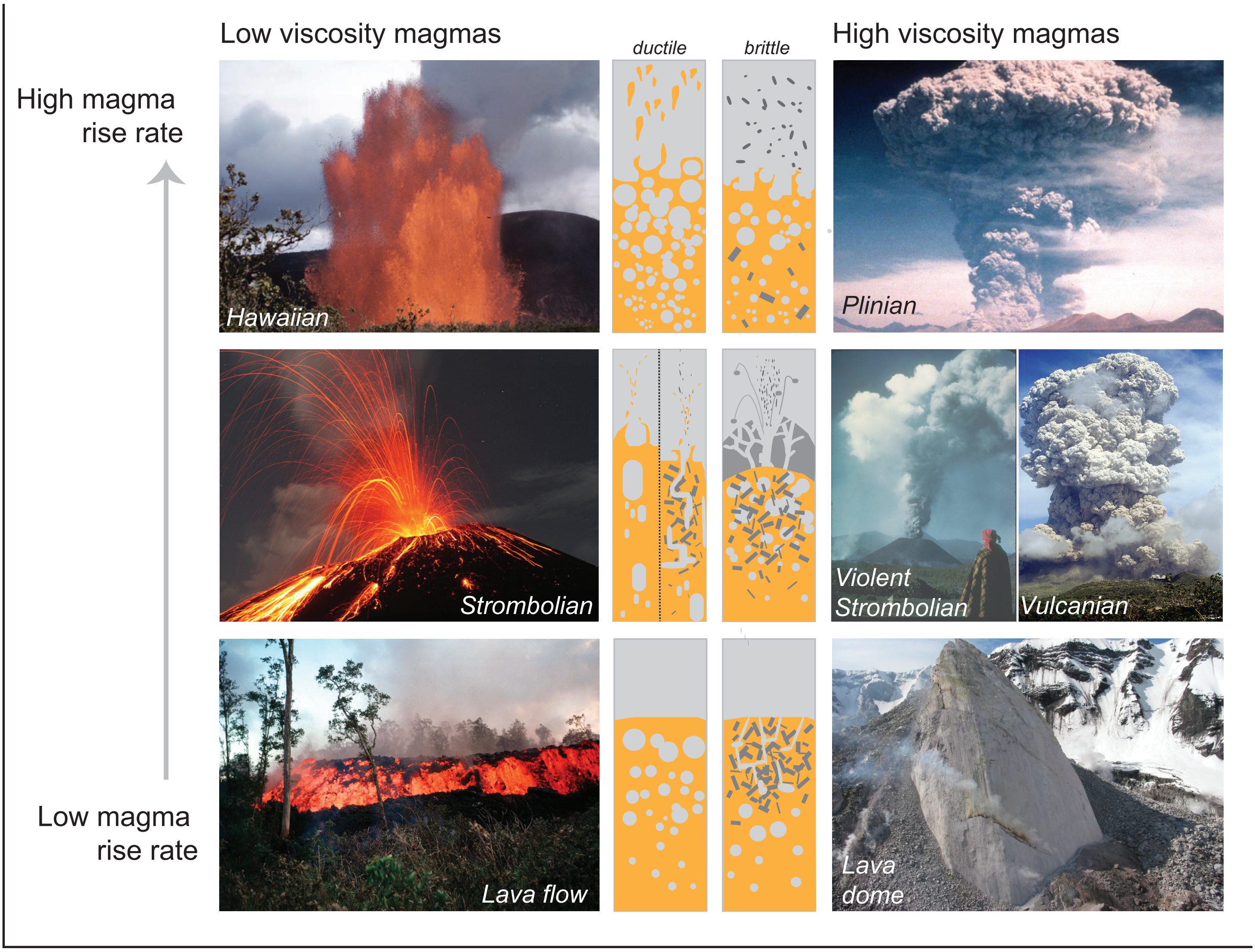
a: Satellite image (AIRS) of a SO_2 cloud associated with the Okmok (2008) eruption



b: An exsolved gas phase in the magma reservoir causes **large sulfur-rich gas clouds** and **muted syn-eruptive ground deformation**




a
Gas-rich eruptions
Liquid
Mush
Magma recharge
Gas accumulation
Gas transport
b
Relative velocity of the exsolved vapor phase
Buoyant bubble rise in liquid
Gas flow through fractures in mush
5%
Gas volume fraction
25%


Low viscosity magmas
ductile
brittle
High viscosity magmas
High magma rise rate
Low magma rise rate
Hawaiian
Plinian
Strombolian
Violent Strombolian
Vulcanian
Lava flow
Lava dome