(Péwé, 1959); at the surface, rims grow wider (from slumping) and taller (from additional volumetric expansion of the wedge) with increasing age (Berg and Black, 1966). There is considerable disagreement whether growing polygon shoulders can ever coalesce and aggrade into high-centered sand-wedge polygons. Berg and Black (1966) report no evidence of sand-wedge polygon shoulders coalescing into high-centered polygons, consistent with observations by Sugden et al. (1995), Marchant et al. (2002), and Marchant and Head (2007) identifying multiple generations of cross-cutting wedges in some sand-wedge and sublimation polygons. In contrast, Sletten et al. (2003) postulate the presence of a closed-cell cryoturbation cycle (see next section) that transfers wedge material into polygon interiors. Lastly, sand-wedge polygons can be distinguished from ice-wedge polygons, as sand-wedge polygons form in hyper-arid regions with dry active layers (Bockheim et al., 2007; Marchant and Head, 2007), meaning that they rarely form with accessory landforms that require seasonal thaw and active layer saturation, such as pingos, alases (thaw lakes), solifluction lobes, thaw caves, or complex fluvial systems (Leffingwell, 1915; Lachenbruch, 1962; Washburn, 1973; French, 1976; Fortier et al., 2007).

Composite-wedge polygons (Fig. 2) represent an intermediate landform between ice-wedge and sand-wedge polygons, and accordingly, display a range of surface morphologies (Berg and Black, 1966; Murton, 1996; Ghysels and Heyse, 2006; Levy et al., 2008a). Although composite-wedge polygons are commonly observed in exposures of inactive, buried permafrost in the Arctic (Murton, 1996; Ghysels and Heyse, 2006), Antarctic composite-wedge polygons are currently forming under present microclimate conditions, providing insight into active composite-wedge polygon development (Berg and Black, 1966; Levy et al., 2008a). Composite-wedge polygons include examples of entirely flat lying microtopography, slightly high-centered polygons, and low-center polygons with gently raised shoulders (Berg and Black, 1966; Levy et al., 2008a). An important aspect of Antarctic composite-wedge polygons is that they form in microenvironments in which seasonally produced meltwater is ephemerally present in locally concentrated, integrated reservoirs, such as snow packs, gullies, and hyporheic zones, providing a range of accessory landforms that can be used to distinguish composite-wedge polygons from sand-wedge or ice-wedge polygons (McKnight et al., 1999; Head et al., 2007a; Levy et al., 2008a).

Active sublimation polygons (Fig. 2) consist of high-centered to domical, mounds of sediment-covered ice, bounded by deep, non-raised-rim troughs (Berg and Black, 1966; Marchant et al., 2002; Marchant and Head, 2007). Sublimation polygons form in the absence of a liquid phase (Marchant et al., 2002). The presence of massive, sublimating ice in the polygon-forming substrate allows sublimation to balance or exceed lithic input to the growing sediment wedge (Marchant et al., 2002). Coupled with the trapping of coarse grains at the fracture surface, complete filling of thermal contraction cracks is minimized, preventing the formation of raised shoulders in sublimation polygons (Berg and Black, 1966; Marchant et al., 2002). Sublimation polygons may form on ice-cored moraines, atop stranded glacier ice, or on debris-covered glaciers (Berg and Black, 1966; Marchant et al., 2002). Sublimation polygon troughs may be oriented by debris-covered glacier flow-related stress (Levy et al., 2006), and commonly develop from fine surface fissures to depressed troughs tens of cm to >1 m deep (Marchant et al., 2002; Kowalewski and Marchant, 2007; Kowalewski, 2009). Surface furrows may be present in active sublimation polygon troughs, indicating the winnowing of fine material into open subsurface fractures (Marchant et al., 2002). More commonly, however, mature troughs are surfaced by gravitationally-slumped boulders, cobbles, or a coarse gravel lag too large to enter the wedge (Marchant et al., 2002) (Fig. 2).

2.4. Resurfacing and polygons

Modification of polygonally-patterned surfaces by polygon-forming processes such as continued fracturing, thermokarst formation, gravitational slumping, sublimation of subsurface ice, etc. all combine to continually change the initial record of thermal contraction crack processes in polar landscapes. Ongoing fracturing and ice melting/sublimation alter the composition and microtopography of polygon interiors, while gravitational slumping and thermokarst formation tend to alter the shape of polygon troughs. Understanding the mechanisms that degraded and rework polygonally-patterned surfaces is critical for assessing the age and duration of polygonally-patterned landscapes, a process complicated by the lack of a reliable chronometer intrinsic to the polygon-forming process (see previous section).

In regions with traditional (saturated) active layers, annual freeze–thaw cycles produce a range of archetypical landforms indicative of reworking of surface soils and rock clasts by ice-related processes (cryoturbation) (Washburn, 1973; French, 1976; Bockheim and Tarnocai, 1998), including frost-heaved/jacked blocks and complexly sorted sediments and rocks. Cryoturbation is a disruptive pedogenic process, and results in mixing of the ground surface by the displacement, rotation, and sorting of sediment particles (Fox, 1994). Cryoturbation is driven by inhomogeneities in the freezing, thawing, and thermal expansion of water in the near-surface (Fox, 1994). Long-term warming and thawing of ice-cemented permafrost in tundra environments generates water-saturated conditions that drive cryoturbation of wet active layers, resulting in the reworking of polygonally-patterned surfaces, such that much of the Earth's seasonally-active permafrost is less than several tens to hundreds of thousands of years old (Gold and Lachenbruch, 1973; Washburn, 1973; Denton et al., 1989, 1993; Yershov, 1998; Sletten et al., 2003) (although recent reports of ice-wedges approaching $\sim$700 ka in age suggest that preservation of deeply buried permafrost, below saturated active layers, may extend through some glacial/interglacial cycles, e.g., Froese et al., 2008). This polewards retreat of terrestrial permafrost environments since the last glacial maximum is strikingly analogous to changes in the distribution of ice stability thought to have occurred due to obliquity variations in martian history (Jakosky and Carr, 1985; Mellon and Jakosky, 1995; Laskar et al., 2002, 2004; Head et al., 2003; Forget et al., 2007; Marchant and Head, 2007).

Traditional cryoturbation is a water- and ice-activated process, and accordingly, it is most strongly observed in poorly drained, silty or peaty soils supporting a wet active layer (Bockheim and Tarnocai, 1998; MacKay, 2000). Cryoturbation in saturated sediments is thought to arise from convective cycling of sediments driven by the propagation of freezing fronts through the active layer (from the surface downwards, and from the ice table upwards) (Vliet-Lanoe, 1991). Convective behavior accounts for the formation of a range of sorted soil and rock structures driven by ice-lensing (Kessler and Werner, 2003). A half-cell of convective cryoturbation behavior has been observed in peat-surfaced ice-wedge polygons (MacKay, 2000). MacKay (2000) shows that the long-term contraction of ice-rich active layer soils (with up to ten times greater ice mass than soil mass) is constrained by the polygon center, while the expansion of the active layer soil is unconstrained by the free surface at the polygon margin, resulting in a net transport of active layer material towards the polygon margin as unmelted permafrost continues to expand through the summer beneath the melting front. Laterally transported material accumulates in raised polygon shoulders or falls into polygon troughs, where it locally increases the height of the permafrost, resulting in the formation of secondary, syngenetic wedges (wedges than develop in a thickening permafrost